QUARTERLY JOURNAL
OF THE
ROYAL METEOROLOGICAL SOCIETY

Editor: J. S. SAWYER

Vol. 87 OCTOBER 1961 No. 374

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PUBLISHED AND SOLD BY THE
ROYAL METEOROLOGICAL SOCIETY
49 CROMWELL ROAD, LONDON, S.W.7
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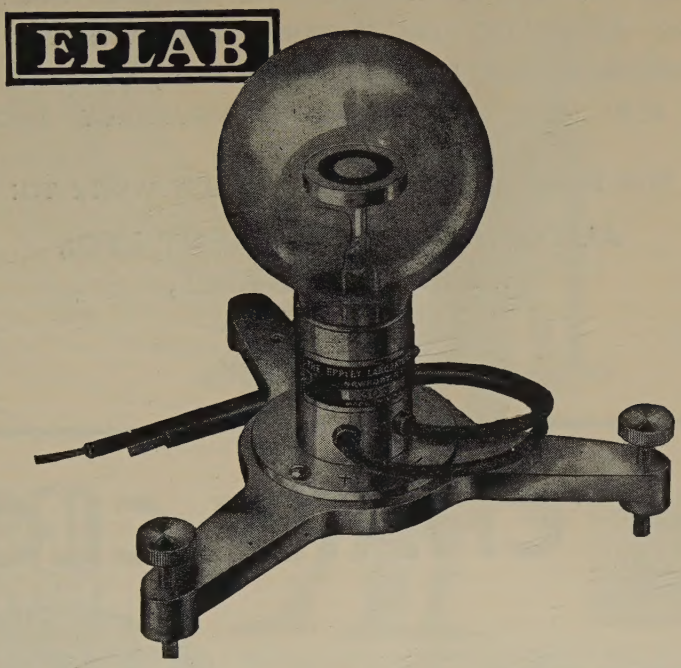

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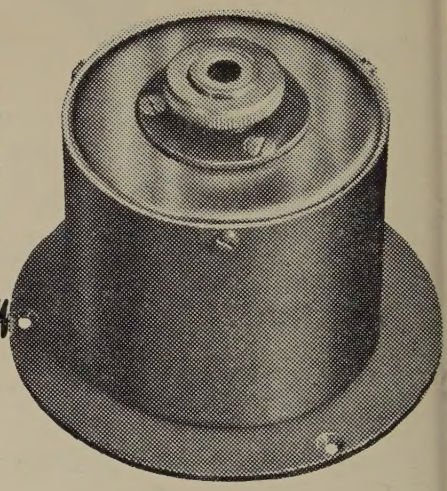

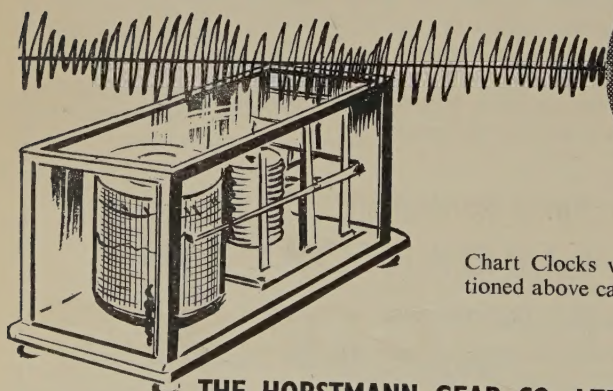

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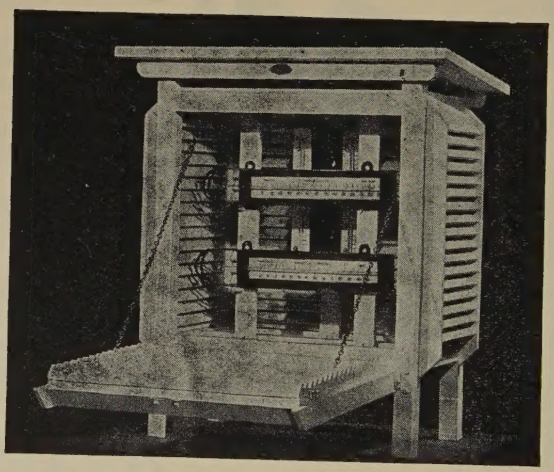

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President of the Royal Meteorological Society 1959-1961

QUARTERLY JOURNAL
OF THE
ROYAL METEOROLOGICAL SOCIETY

Vol. 87 OCTOBER 1961 No. 374

551.509 (04) : 599.9

Meteorology and the community

By J. M. STAGG, C.B., O.B.E., M.A., D.Sc.

Presidential Address delivered before the Royal Meteorological Society on 19 April 1961

In recent years it has become a pattern for the two addresses which it is the privilege of the President to give that the first is devoted to a scientific theme and the second to a more general survey, and I had planned to adopt this arrangement, though in the reverse order. I had looked to release from official duties to provide opportunity for completing one or other of the lines of investigational work I had long been interested in so that I might present the results this evening. Perhaps the motives were not wholly so pure and altruistic as this sounds : there was also the idea of getting them published under cover of a Presidential Address, if you do me that honour, without having to run the risk of rejection by the Editor. But regrettably for me the work has not been so accommodating : the results don't seem to fit in with current theory or the theory with my results, and I cannot make up my mind whether it is my results that are at fault, or the current notions that are awry. So they must await another occasion.

In the meantime, my recently found freedom from gainful employment has indirectly provided an alternative theme : for by allowing me to extend the circle of meetings with members of the lay community, including among them those whose public views about the national meteorological service were formerly liable to be coloured by their knowledge of my profession, this new latitude has led me to reconsider from the layman's angle the contact points of meteorology with the public. It has also provided the opportunity of contributing, as I believe, to the number of those who now use the national weather service more frequently and intelligently than they have hitherto used it, but that is incidental to the present purpose. The impact of meteorology on the public is no new theme for the Society : whenever else, it was discussed at the centenary celebrations in 1951. But the ten years since then have seen so many important changes in the science and in the means of communicating its products to the public, that it probably deserves another airing. Besides, although about one-tenth of our membership, according to some random pages from the new List, are members of the national meteorological service, and know of the continuous efforts that are made to satisfy public needs, the other nine-tenths are less well-informed. Anyway, we are all members of the public both in a supporting and a consuming capacity and the Society is the nearest approach to a consumers' association that we have in this country.

I shall be concerned almost wholly with weather forecasting : not because the impact of climatology is less impressive than forecasting, but because it is of a different nature. Through the advice for planning (in the broad sense) it can give for almost every human activity, climatology makes contributions of value to the community out of all proportion to the cost of the service. But though more pervasive and enduring in its benefits, climatology's impact is made through fewer customer channels, and is therefore more easily controllable. For every hundred members of the community who in the course of a year use planning information and guidance based on past weather, there are probably millions who have direct contact with forecasts. It may well be that even now only a small fraction of the use that could be made of the vast accumulation of climatological data that are stored up and down the world has so far been made, partly because of the time and effort hitherto needed for dealing with the material, and also partly because many, or most, of the potential users are still unaware of its value. The first is rapidly

being remedied by the introduction of speedier processes for handling data of this kind, and the second is mainly a matter of priority in the allocation of resources, particularly in manpower for public relations and salesmanship.

With forecasting, the position is quite other. For over a century forecasts have been distributed as widely as the prevailing media have allowed. Except in war time they have been broadcast by sound radio for nearly 40 years, by television for 12 years and for six of those by the forecaster himself. So that by radio, television or newspaper, everyone in the country now has access to several forecasts a day, and there is no question that many set great store by them. But is the proportion of the population which does appreciate them as great as we sometimes assume, and does even this proportion value the forecasts as highly as they deserve ? In an age when an article or a service is esteemed at the price paid for it, value is perhaps not the right word. In this country the cost of the whole meteorological service is around 1s. 6d. per annum per head of the population and that 1s. 6d. includes many things not directly connected with forecasting. As there is little one can get several times a day throughout the year for that kind of price, could cheapness and too ready accessibility be reasons why weather forecasts do not command the wide respect that might be thought appropriate to guidance which has so much organization and scientific thought behind it ? Probably not : for leaving aside groups like air-crews, both military and civil, for whom specialized services with facilities for direct contact between the meteorologist and the customer are provided at a charge to the department concerned, there are many sections within the community who show their faith in weather forecasting by going out of their way to supplement and improve on the advice they get from broadcasts and press by telephoning to a forecast office or by using the automatic telephone weather service, and increasingly large numbers of the population use these services. But is there not, even here, a pointer to some weakness ? For even allowing that the G.P.O. has provided only ten or a dozen cities with facilities for an automatic service and allowing that the cost of telephoning to the nearest forecast office may be above the local minimum rate for a call, yet a single reliable forecast can so far outweigh the inconvenience and cost of telephoning that it comes as a surprise to learn that in such a weather-conscious country as ours, the total of inquiries in both categories is running at five or six millions per annum, which is little more than one call a year by one in every ten of the whole population, or one call every other year by each family in the country. As most of the calls are probably made by people who have already found that they can get by this means the kind of service they need, the fraction of the population who frequently supplement the broadcast service in this way is almost certainly even smaller and though it is increasing each year, the rate of increase is hardly such as to point to its ever being likely to be more than a small minority.

So it comes to this : that if we discount the press as a not very influential medium for disseminating forecasts, we seem to be led to conclude that broadcasts of weather forecasts by sound alone or by TV are the main channel by which most people have contact with our science. And when we note that there are now more than twice as many licences for TV as for sound alone we can hardly escape the conclusion of the current tag that for most people ' meteorology is the weather we see on the telly.'

Now there was ample evidence that dissemination by this medium at first quickened and widened interest in weather forecasting and in its potential usefulness, especially after personal presentation was introduced. It is therefore the more noticeable that though the daily quota of TV programme hours has at least doubled since the B.B.C. service began, the time allotted to personal presentation has been cut from five to three minutes a day. If viewers value the service why, we may ask, have they not resisted this reduction or even demanded more detailed and more frequent presentations ? In other departments and categories of interest the broadcasting authorities have ways of sensing what the viewer is believed to want, and they find ways of providing it. Can it be that the surge of interest in the early days stemmed primarily from the novelty of the presentation as entertainment, and that viewers have not learned to value the content progressively more as the novelty has worn thin, and if so, what may be the reason ? One explanation could

be that the lay viewer, untrained in thinking in terms of television weather charts with their daily transformations of segments of pressure patterns, may not have enough time during a high-speed concentrated presentation to take in all that he sees and hears, or to consider adequately the implications of the broadcast for his own purposes : as he has only partially comprehended what was said and seen, he is also unable to retain his impressions clearly enough through the items of subsequent entertainment. Another explanation, which could reinforce the first, might be that in too many instances the viewer's confidence in the average day-to-day dependability of the forecasts is below the threshold of confidence that he needs to make the effort to remember the implications, and as he has no means of gauging whether any particular forecast is likely to be more or less accurate than the average, he gets no more and no less out of a good forecast than he does from one that is not so good.

The first of these two explanations could be related to the mechanics of presentation, to the distractions on either side of the forecast in the programme sequence, and more especially to the short time allowed for it. Those are obvious factors which can and probably do in some measure detract from the value the viewer derives : but, within limits, they are probably not quite as decisive as the frame of mind which the viewer brings to the forecasts. With confidence in them his concentration is sharpened, and he will get out of them what he needs; and contrariwise, reduced confidence dulls his receptivity and with it his appreciation. The extent and degree of this lack of confidence are matters of opinion, but accepting that an element of it does exist, we might consider whether anything simple can be done to reduce it.

It would serve little purpose, even though it were probably true, to tell the layman that the service is much more accurate and reliable than he thinks. If a numerical measure were given, he might reply with a reference to Shaw's 1911 text-book on the subject : this describes how the forecasts issued to the press at that early period of meteorological history were regularly checked in respect of force and direction of the wind, state of sky, rain, temperature and other phenomena such as frost, fog, and thunder; the checks showed that 87 per cent of the forecasts gave a fairly accurate indication of the weather for the forecast period. This percentage would require explanation against whatever was given as the current figure. The layman, could, of course, be told that assessment of accuracy is a more difficult task than making the forecast and also that accuracy is very much a matter of how it is measured and who does it : without a generally accepted, practical and readily understood criterion there can be room for much difference of view. The accuracy of the forecast as the layman receives and interprets it can be quite other than the forecaster's assessment of what he intended it should convey and how it should be used; and there is the accuracy as it might be computed by a wholly disinterested adjudicator. As big a disparity could exist between any pair of these assessments as between any other pair. Figures of merit do not take us far : something more is needed. If we accept that the best way of getting a forecast is by personal consultation with the forecaster, part of that *something* should go some of the way to make up for lack of contact between the viewer or listener and the forecaster. Now one advantage of consultation is that the customer can discuss alternative possibilities and another is that he can get some indication of the forecaster's confidence in his advice. To discuss alternative possibilities of synoptic development in a broadcast presentation is not wholly impracticable : it is sometimes done and should probably be done more frequently, even if it requires a minute more time. And to give an indication of the forecaster's confidence in the line he selects could be a regular feature. The forecaster has long had his own code for expressing the degree of likelihood of occurrence of particular elements in his forecast, but it is doubtful whether the niceties of gradation from unqualified futurity to ' chance of ' are not swamped in the listener's mind by the other technical details in a highly compressed statement often made at high speed. So whilst retaining this code for those who can use it, there might be merit in introducing an additional simple index of the forecaster's confidence in the main sections of his forecast – A for high confidence, B average, and C low. Besides serving to remind listeners of the difference between a

weather forecast and a rigorous prediction, this could guide them to a more intelligent use and appreciation of the forecasts. It would be consolatory and also heartening to overhear one layman reply to another who had been grumbling about being let down: 'But that was only a C forecast, so you should have known to be wary of what it said.'

Returning to the wider aspects of the relations between organized meteorology and its customers in the community, there are good grounds for believing that the man in the street, and certainly the man in the country, probably has a keener weather sense than we meteorologists sometimes give him credit for. He knows well enough that weather is apt to run in types, and for the most part he plans his day accordingly, using the forecast for confirmation. For this reason a proportion of the forecasts he gets – of the kind that tell him tomorrow will be much as today – do not much impress him with the triumphs and potentialities of meteorology as a science. Nor perhaps do the relatively greater detail and accuracy of shorter period (6 and 12-hour) forecasts. In this fast-moving era of motor cars and aeroplanes it is not an uncommon experience to race the weather across country; so the layman may be excused for thinking that a frequently reporting network, perhaps with one or two storm-warning radar equipments in it, can on most occasions provide a sound enough basis for extrapolating over the next few hours, without the elaboration of a complex science behind it. What is more liable to determine his confidence in forecasting and his respect for meteorology is a reasonably frequent experience of accurate guidance, given at least 12 hours in advance, for those days on which the weather turns out to be different from what he, by his own native wit and experience, would have expected. Advice for short periods is valuable : for that there is the evidence of the expanding use of the automatic telephone weather system. But as meteorologists, we have to admit that this need is met less by meteorological science than by reliable observations and communications, together perhaps with some elementary rules that have stood the test of time.

Nor is the layman much appeased by evidence that meteorology advances steadily, from year to year. The kinds of benefit to be derived from ocean weather ships, sferics networks, radar, the international geophysical year, electronic machines, rockets and satellites, have each in their turn been well documented. Unfortunately these benefits to the science come to be too often linked in the public mind with the promise of improvements in forecasting, so that every new aid carries with it the prospect of greater reliability in forecasts. But for the kind of improvement the layman looks for the prospects remain very much as they were.

So far I have analysed what I believe to be some of the views of a cross-section of the public, or, if you like, the views of representative members of this Society who have not had the privilege of serving in the Meteorological Office. From now on my own views may here and there obtrude more obviously.

It is, for example, giving away no secrets to say that no one is more cognisant of the need for improving the reliability of the forecast service than the forecaster himself, and those responsible for his work. Improvement is, and always has been, a prime preoccupation. At various times through the years, the spotlight of enquiry within the service has ranged over every stage of the process : the content, times and frequency of the reports and the density of the reporting network, the organization for collecting and handling the material, the technical procedures and methods of thought of the forecasters and the ancillary aids at their disposal, the contributions made by research in improving basic knowledge and theory, the methods of presentation and checking of the forecasts, the number and quality of the forecasting teams and the supervision of the whole process – all these have been examined and modified repeatedly. Yet if we are concerned, as I am here, not so much with specialized aviation forecasts based on the most recently observed winds and temperatures, or with details for short periods in some particular locality, but with the general 24 or 30-hour forecast on which plans for work-a-day or holiday activities are made in the evening for tomorrow's daylight hours – for this kind of forecast it would be hard to substantiate claims for an accuracy as high as Shaw claimed for his forecasts 50 years ago. Yet since that time the science, the techniques of application, the apparatus

and the organization have developed almost out of recognition. Where is the answer? Can it possibly be that forecasting in this country has in some way become tougher, or did the early practitioners have something, some special know-how or experience, that their successors nowadays haven't got? The forecast problem in every synoptic situation can be put into one or other of several categories which range in difficulty from the comparatively straightforward to the almost impossibly complex. I have come to think of four categories A to D defined by the kind of skill and experience needed to make reasonably successful forecasts :

A – Average competence and elementary rules	55% ± 10
B – Real technical skill and experience	25% ± 5
C – High scientific skill, with real insight and long experience	15% ± 3
D – Intractable : exceptional scientific skill and experience, or luck	5% ± 1

The last category includes those situations in which luck plays as large a part as anything else, either for want of information from critical areas and perhaps also the theory to apply to it, or simply because of brute instability.

The number of categories and certainly the percentage of cases in each category indicated in the column on the right are personal assessments, and they vary from time to time. In particular it is, I believe, the quite random variations in frequency of the last two categories, C and D, of very difficult and unpredictable cases which lead to the perplexing feature we all recognise. For weeks or even months together forecasts continue at a high and consistent level of reliability, as if theory and practice had made a breakthrough : then, just as this level of accuracy is becoming accepted as a permanent new standard it reverts once more. In a similar way, it could be that Shaw's checks referred to a synoptically easier period, before the warming up of this part of the world had progressed far and therefore with fewer changes of type or less variability of weather from day to day. This is an unlikely explanation; and it is equally unlikely that the early forecasters had access to any special skill, apart perhaps from that derived from great length and continuity of experience at their simpler job. It is also certain that fluctuations of standard nowadays are not symptomatic of varying effort or diligence. On the contrary, the greatest vigilance has constantly to be maintained to keep the level of accuracy as high as it is. It may be paradoxical but it could also be true that a greater effort has to be made now, than formerly, merely to uphold a similar reliability for the 24-hour forecast.

What I have in mind here is the influence on the quality of the main forecasts of some recent trends, particularly two, namely the ease and rapidity of communications coupled with the frequency of reports, and the demands from broadcasting and press authorities for more and more bulletins and snippets of forecast to accommodate their programme schedules. Before the introduction of 3-hourly and hourly charts the forecaster had to extract all he could from every bit of information and put all he knew – even though it may not have been much – into his diagnosis. Nowadays reports keep pace with the weather from hour to hour, and though they are probably a necessity for specialized purposes at some forecasting offices, the fact that they exist at all means that the central forecaster, too, can ignore them at his peril. This, along with the distractions that come from keeping up-to-date with the demands of broadcast and other schedules, has dubious advantages. With each new bulletin, it is true, he has opportunity of adjusting his forecast to keep it in phase with the weather and with the immediate short-period prospects : but a proliferation of forecasts can cause confusion at the receiving end, and perhaps more seriously, the obligation to keep abreast of the stream of reports can lead to modification of the main issues when no change might have been a sounder practice.

But this is well enough understood, and anyway, forecasters have enough to contend against without carping from a Presidential Address. Indeed, I would not have ventured along this rather censorious line of reflection were they not now assured that they probably never had brighter prospects ahead of them. For there can be little doubt that the introduction of machine methods into the forecasting room has affected the whole future outlook. As ages go in forecasting practice these methods are very young : even

in larger-scale analysis and prognosis perhaps only their potentialities have so far bee demonstrated, and much still remains to be done before they completely replace conver tional methods in this part of the world. It is, for example, recognized that in the endles procession and variety of behaviour patterns in the atmosphere there are incidents whic have much significance for tomorrow's weather but are below the threshold which th numerical grid can grasp : and, apart from these, it has also to be shown that there d not exist grades of difficulty for the machine as there do for the methods hitherto usec It would be a happy and convenient division of labour if the forecaster's most vexatiou cases (C and D categories on p. 469) turned out to be those that could be reliably le to the machine : but even if the opposite were true, so that the present, if temporary ceiling in the machine's capabilities had to be interpreted as meaning that the new aid can deal most successfully with those cases (A and B) that the forecaster himself find least troublesome with his present methods – even then the immediate advantage to th forecasting service would be significant. For if we assume that the forecaster will be abl to distinguish the occasions when the machine's opinion has to be treated with respec from those when it can be ignored with impunity, he can reserve his greatest efforts fo the most difficult situations. Believing that two brains are generally better than one, have long wished it were practicable for the Central Forecasting Office to have two team of equal skill and experience working independently and as far as practicable using differen methods. For obvious reasons that was an unattainable luxury, but the machine now provides the second opinion, with the added advantage that it works to clearly prescribec rules and assumptions. So the upshot is that the forecaster will have his confidence in creased when the two opinions converge, and this can be reflected in the clarity of his statement; and when there is disagreement he will be able to concentrate on the factors known to be omitted from the numerical rules and so even then the sector of doubt will be reduced.

Meanwhile, as we learned at the December meeting of the Society, much thought is going into automatic processes for making, recording and collecting weather reports. In the climatologist's work for the public, any developments which make it easier for him to get access to his data and deal more speedily with them are good things. With the reservations of my earlier argument this is also true for the forecaster and he may derive an added bonus. For it happens that the observational elements which can be recorded and communicated most economically by automatic means, (pressure, temperature, humidity, and wind) include the primary elements used in numerical prediction. So when the reporting of those elements from a sufficient network of stations becomes fully automatic the machine can proceed with producing a set of first-approximation charts, ready to take the other elements such as the type, amount and change of cloud and present and past weather, which require human judgment, and from which the second-order refinements of analysis and prognosis are derived.

Not all mechanization necessarily conduces to real improvement in the finished product but it looks as if current trends in the mechanization of weather forecasting practices could be beneficial for the forecasts besides relieving the forecaster of some of his more routine tasks. A possible exception might be the premature extension on too widespread a scale of so-called objective methods of dealing with specific weather elements such as the future state of cloudiness or fog or upper wind. In contrast with numerical prediction of the large-scale fields of pressure and contour height which aims at using basic principles even though only approximately, objective methods employ statistical relationships that have been found to exist over a particular period between two or more measurable parameters and the element to be forecast. These objective methods have their place. In the years ahead when the whole forecasting process is dehumanized it may be necessary to accept procedures of this kind for clothing the mechanically predicted contours with mechanically forecast weather, and so it is justifiable to be experimenting now with these auxiliary feeder lines of the future. At the training stage and in times of rapid dilution of forecasting experience it may also be expedient, though sometimes dangerous, for the neophyte to have another assessment to appeal to. But even though there were authentic

rroboration that these correlation methods are as reliable as the skilled and experienced ain, not just in the average case but in critical circumstances when the atmosphere is having as it so often does in a quite unexpected and hitherto unrecorded way, and when uch is at stake – even then it is likely that forecasting will continue to benefit by being ought about and practised, at least until it is within sight of being completely mechanized.

Increasing mechanization is also likely to have its effect on the functioning of the entral Forecasting Office and its relations with the public, both direct through the isseminating agencies and through the intermediate auxiliary offices. It has always been ebatable whether the central office should work in isolation, concentrating wholly on nalysis and large-scale prognosis, or whether it should carry the process through to the etails of weather, at least for the larger regional areas. In this country a compromise as been favoured by which the customer has limited access to the central office, on the round that it is salutary for the forecaster to learn at first hand, from time to time, what is professional colleagues (who serve as his wholesale or retail agents at the outstations) nd what the lay public think of him and his works. But a greater measure of disengage- ient may now be forced on him. Already the output of the high-speed automatic teletype sed in the international exchange of reports confronts the central forecaster with ever- ncreasing amounts of data : if all these data are really relevant to the production of orecasts for the next day or two the forecaster will require complete freedom from dis- raction for studying them. The need to concentrate in one office the expensive equipment ssociated with numerical prediction and to conserve the limited number of forecasters who ave the scientific skill, retentive experience and flair necessary for maintaining a successful eam may also require the central office to shed everything but basic analysis and genuine orecasting, leading to a greater differentiation of function between it and the outstation ffices. Apart from being unable to get all the basic information that goes into the central mporium without duplication of costly high-speed communication systems, these offices ave exacting enough responsibilities in concentrating on the short-term prospects in heir area with the help of storm radar and similar technical aids, and also in tailoring nd packaging the centrally produced guidance, not so much by using the techniques of orecasting as by applying meteorological physics and regional experience. This clearer dichotomy of function presupposes two-way confidence between the central and subsidiary ffices; but this is now less of a problem than it was inclined to be when both categories f office were staffed from the same scientific class.

There is time for only a brief reference to a more drastic possibility in dissociating the central forecasting function from the various outstation organizations which are needed for distributing the forecasts to different classes of user and for meeting their particular requirements. It can be contended that the processes of bulk manufacture, comprising all the stages from making and collecting the observational raw material to the preparation of the general analysis and prognosis, are appropriately the responsibility of a state department or a public corporation. Particular public and military services and groups and sectional interests within the community could establish their own channels for distribution from the central office of manufacture through wholesale and retail agents to their individual customers. And, in a similar way, an organization specifically designed to meet the needs of the public would look to the Central Forecasting Office for its broad guidance, and it would work through regional and local retail offices in collaboration with the radio and television authorities and the G.P.O. much as at present. But I do not intend at this late stage to examine the balance sheet of pros and cons for a set-up of this kind or to ventilate the role of private exploitation in this or any intermediate stage of denationalization. It may even be doubted whether it is excusable to raise questions of this kind at all. On balance I believe it is; for though we discuss at our meetings many isolated contributions to the science, we have little opportunity of considering where all the contributions lead and how far progress in the science is reflected in improved service. So we perhaps owe it to ourselves to make opportunity from time to time to think around these aspects both as members of the Society who are deeply interested in the prestige of the science and also as members of the public who use its applications.

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Winds in the upper atmosphere

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(Manuscript received 7 December 1960. Communicated by Prof. P. A. Sheppard)

SUMMARY

A systematic study of the winds in the upper atmosphere has been carried out during the 5-yr per 1953 to 1958. At heights of 85 to 100 km regular prevailing and periodic tidal wind components have b resolved. Components of the prevailing wind of about 20 m sec^{-1} towards east are observed during the sumn and winter months, with similar westward components during spring and weak westward winds in autu Large southward components of approximately 20 m sec^{-1} are observed during the summer. The amplitu of the semi-diurnal periodic wind is also about 20 m sec^{-1}, and this component can usually be represen by a vector which rotates in a clockwise direction with constant amplitude twice every 24 hours. La departures from this simple model occur at certain times of the year. Measurements of the 24-hr and 8 periodic wind systems are also described, and an upper limit of 2 m sec^{-1} is determined for the lunar se diurnal wind component.

1. INTRODUCTION

A systematic study of the wind at heights of 85-100 km above the earth's surfa has been carried out at Jodrell Bank (53N, 2°W) between September 1953 and the en of the International Geophysical Year in 1958. Wind measurements were made b observation of the drifts of ionized meteor trails using high-powered radio-echo equipmen The technique and method of analysis have been described by Greenhow (1954). Th method enables mean wind speeds in two directions at right-angles to be determined f each hourly period of a single day. Subsequent harmonic analysis enables the variatio of the velocity during any period of 24 hours to be resolved into prevailing (constant diurnal (24-hr), semi-diurnal (12-hr) and higher periodic components. Observatio

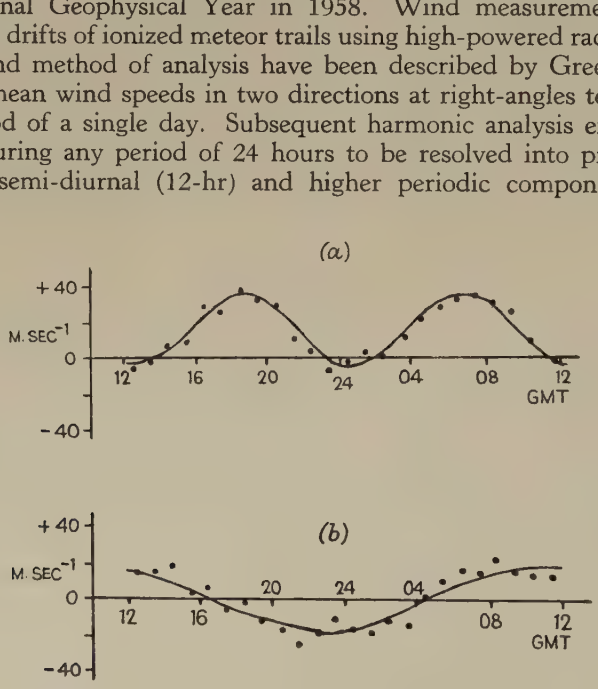

Figure 1. Examples of the results showing the hour-by-hour variations of the wind throughout the da The curves show the sum of the first three terms of the fourier series.

(a) The east-west component for a typical September day
$$V_{W\to E} = +16{\cdot}0 + 1{\cdot}9 \sin(\pi/12\,(t - 6{\cdot}3)) + 19{\cdot}8 \sin(\pi/6\,(t - 3{\cdot}7))$$

(b) The north-south component for an April day
$$V_{S\to N} = +1{\cdot}1 + 19{\cdot}3 \sin(\pi/12\,(t - 4{\cdot}8)) + 3{\cdot}0 \sin(\pi/6\,(t - 2{\cdot}5))$$

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e usually divided into hourly intervals, and some examples of single days' observations e given in Fig. 1. North-south and east-west components of the wind are shown for two ays in September (a) and April (b). In Fig. 1 (a), the wind is clearly composed of a eady drift of about 20 m sec^{-1} towards east, together with a large semi-diurnal variation f 20 m sec^{-1}. Harmonic analysis shows that the amplitude of the 24-hr component is nly about 2 m sec^{-1}, which is hardly significant on this particular day.

The April day in Fig. 1 (b), on the other hand, has been selected to show the quite rge diurnal wind component which can at times exist. Because of the seasonal variations oth the north-south component of the prevailing wind and the semi-diurnal wind are nall at this time. The diurnal periodic wind variation is clearly visible with an amplitude f 20 m sec^{-1}. This is an exceptionally high value, and usually the amplitude of the iurnal wind component on any individual day is in the range 0 to 10 m sec^{-1}. This paper s concerned with the diurnal, seasonal and height variations of the various wind omponents.

The ionospheric convention, in which the wind blows *towards* the direction indicated, s used throughout the paper. The overall wind vector is frequently resolved in north-outh and east-west components when a positive amplitude refers to a wind component lowing towards north and east, respectively.

2. Results

(a) *The prevailing wind*

i) *Seasonal variations of the wind vector*

The variation of the prevailing wind with season shows a definite pattern which is epeated in its broad outlines each year. This annual variation is shown in Fig. 2 where he average monthly velocities for 1953-8 are plotted on a polar display. It can be seen hat the highest wind speeds occur during the summer months June, July and August, when values of 15-20 m sec^{-1} towards south-east are present. By September the wind nas backed to practically due east. There is a decrease in velocity in October when the speed is only 2 m sec^{-1} towards north-west before strong eastward velocities return in December, January and February. In Spring, there is a westward excursion of the wind vector before the large south-eastward velocities of summer are again reached.

The separate east-west and north-south components of the vector for each year will now be considered in greater detail.

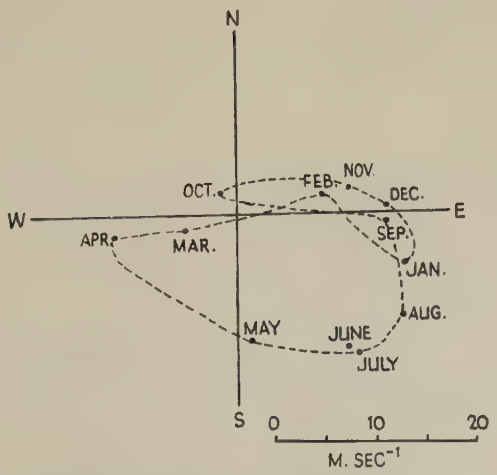

Figure 2. Monthly variations of the prevailing wind vector. Average values for 1953 to 1958.

(ii) The east-west component

The individual monthly values of the east-west component are plotted in Fig. 3 for each of the five years. It can be seen that every year there are strong eastward componen present during the summer and winter months. There is a westward component pres during the spring and a tendency for the wind to go westward during the autumn. T general features of the variation in each year are very similar.

The westward values do, however, vary somewhat both in magnitude and time occurrence. For instance, in April 1956, the wind speed is 18 m sec^{-1} to west, where in the previous year it was only 3 m sec^{-1} at that time. Similarly, in 1955, the westwa component did not occur until April, whilst in 1956 it was already present in Februa

There are even more marked variations in the behaviour of this component in autum In 1953, 1956, 1957 and 1958 there was a definite westward component, whilst this w not so in 1954 and 1955. However, if we examine 1954 in greater detail we find th although the wind was towards east in October at the mean height of observation, the wi direction reversed above 100 km (Greenhow and Neufeld 1956).

The mean seasonal variation averaged over five years is plotted in Fig. 5 (a). T curve shows a semi-annual variation with maximum velocities to east at the solstic Strong westward winds occur in spring and weak westward winds in autumn.

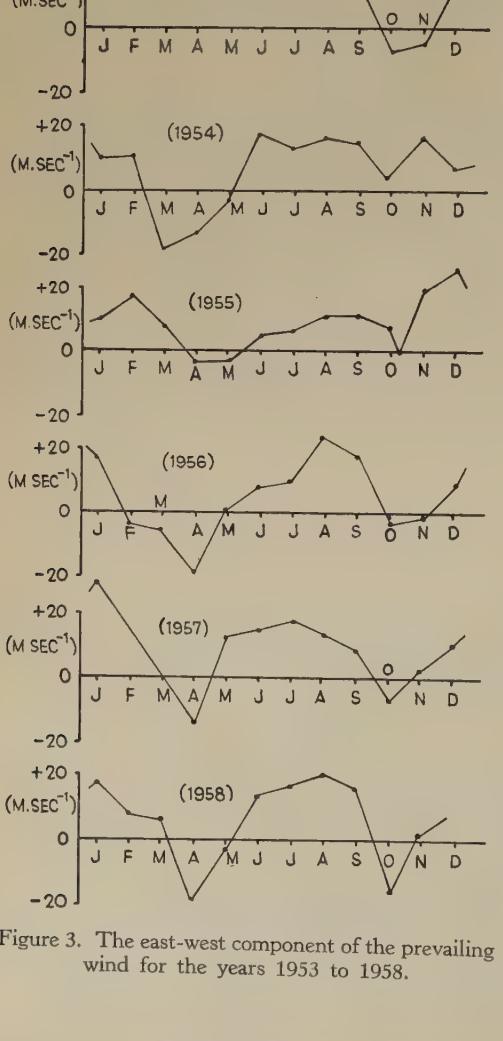

Figure 3. The east-west component of the prevailing wind for the years 1953 to 1958.

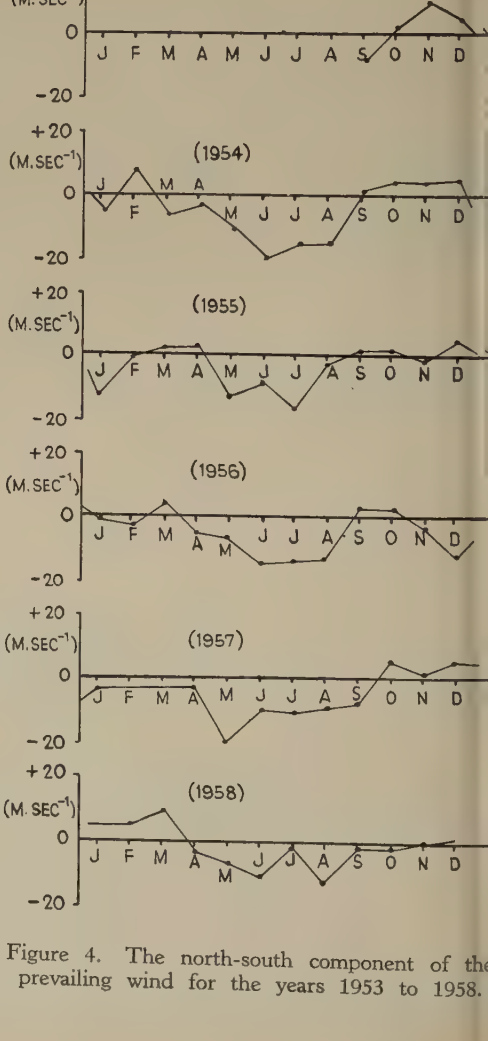

Figure 4. The north-south component of the prevailing wind for the years 1953 to 1958.

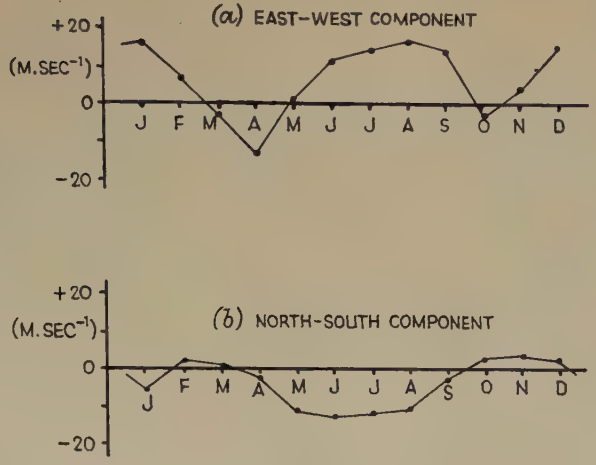

Figure 5. Average monthly values for the north-south and east-west components of the prevailing wind.

(iii) *North-south component*

The monthly values of the north-south component for each year are plotted in Fig. 4. The most obvious feature of the variation is, once again, the similarity between the 5-yr observations. Especially evident are the large components towards south during the summer months. During the rest of the year the north-south component is smaller and more irregular with a small northward component during most of the winter.

The mean curve for the five years, shown in Fig. 5 (b), has smoothed out most of these irregularities and shows that apart from the strong southward component from May to August, the velocity during the rest of the year is small. The monthly values for the north-south and east-west components of the prevailing wind, averaged over the five years, are listed in Table 1.

TABLE 1. NORTH-SOUTH AND EAST-WEST COMPONENTS OF THE PREVAILING WIND

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
Component towards north	− 5	+ 2·5	+ 1·0	− 2·0	− 11·0	− 12·5	− 11·5	− 10·0	− 2·5	+ 3·0	+ 3·5	+ 2·5
Component towards east	+ 16·0	+ 7·0	− 3·0	− 13·0	+ 1·5	+ 11·5	+ 14·5	+ 16·5	+ 14·0	− 3·0	+ 4·5	+ 15·0

(Monthly averages for the 5-yr period 1953-1958. Amplitudes m sec^{-1}; positive towards north and east).

(iv) *Height variations*

During the year 1954-1955, the variation of the prevailing wind with height was measured, in addition to the seasonal variation at a mean height of 92 km (Greenhow and Neufeld 1956). The results for the east-west component are summarized in Fig. 6 (a). During December, January, June and July the wind is towards east with a positive gradient of about 0·8 m sec^{-1} km^{-1}. Extrapolation of the results to lower heights suggests that the wind would become westward below 80 km. During October, November, February and March the wind is again towards east between 80 and 100 km but with a negative gradient (decreasing eastward velocity with increasing height) of 0·5 m sec^{-1} km^{-1}. There is no significant wind gradient during August and September when the wind is towards east, or in April and May when westward components are observed.

In the north-south direction, gradients of about 0·3 m sec^{-1} km^{-1} were present most of the time, with larger northward amplitudes at greater heights.

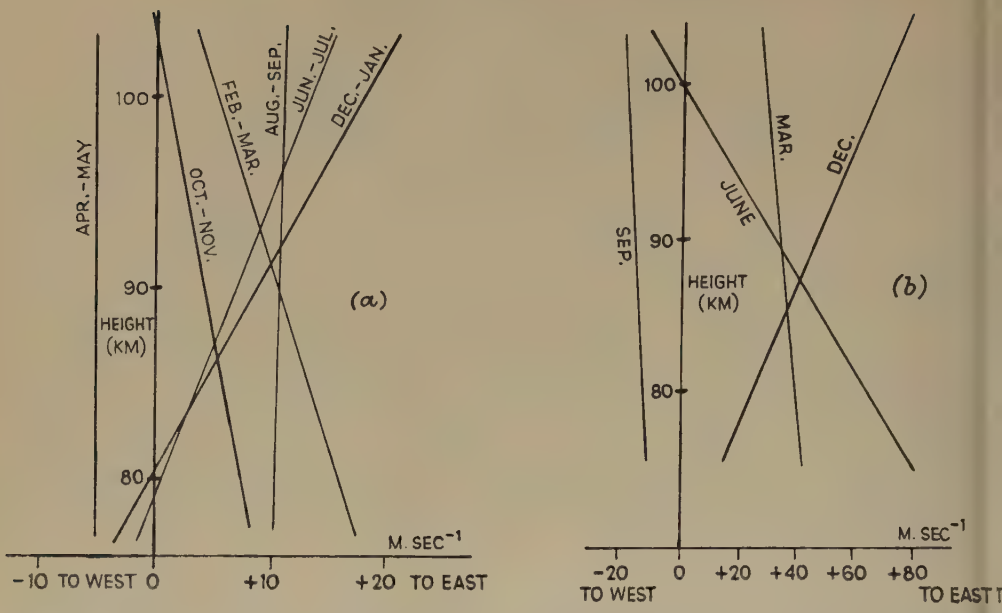

Figure 6. The height variation of the prevailing wind. (a) Manchester observations; (b) Adelaide observations

(v) *Irregular variations in the prevailing wind*

Although the mean monthly values of the prevailing wind show regular seasonal variations which repeat from year to year, the wind may show large irregular variations with periods of several days. An example is given in Fig. 7, where the variations of the wind vector for a 14-day period are illustrated. The wind velocity departs considerably both in magnitude and direction from the mean seasonal value, which is also shown. Variations of this sort have been identified with large pressure systems at heights near 90 km similar in scale to the weather systems at ground level (Greenhow and Neufeld 1960).

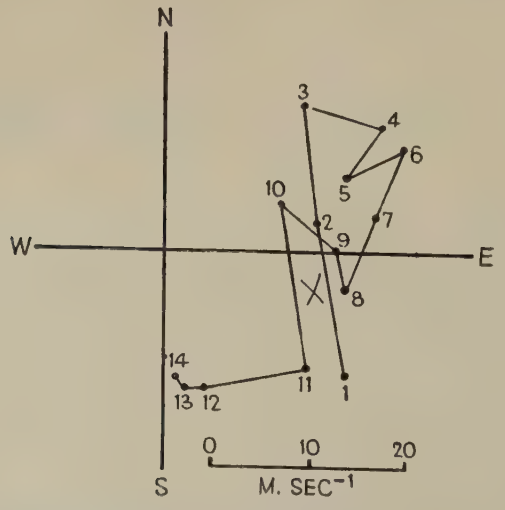

Figure 7. Vector diagram showing irregular variations in the wind over a 14-day period. X = seasonal value for the period.

(b) *The semi-diurnal wind*

i) *Comparison of the 5-yr observations*

The semi-diurnal periodic wind component can usually be determined from the harmonic analysis of a single day's observations, and is given by the second harmonic of the fourier series. It is found that there are large day-to-day, seasonal, and year-to-year variations in the amplitude and phase of the 12-hr wind vector, and this section is mainly concerned with these variations.

It is convenient to summarize the results in the form of harmonic dials, in which any point describes the amplitude of the periodic component, and the time of maximum velocity towards north or east. A series of harmonic dials for each of the five years is given in Fig. 8. To simplify the presentation, results are grouped into a number of periods during which the character of the wind does not show any large systematic changes. The observations during winter (December, January, February), spring (March, April, May) and summer (June, July, August) form three fairly well-defined groups. During the autumn months, however, the phase of the wind undergoes a rapid variation of large amplitude. The points for September, October and November are therefore shown separately.

Consider first the north-south component. In winter, maximum velocity towards north occurs about 0630 (and 1830) GMT. The results for all five years are similar, although there are variations in phase of about 1 hr (30°) from one year to another. The amplitude of the wind component also varies appreciably, being 30 m sec^{-1} in 1953-4, but only 10 m sec^{-1} in 1955-6. During spring and summer the amplitude of the semi-diurnal wind is smaller than in winter, being of the order of 10 m sec^{-1} in spring and 15 m sec^{-1} in summer. The phase of the oscillation is also earlier, maximum velocity towards north occurring between 0430 and 0600 hr. The phase in summer usually leads that in spring. Once again variations of about 30° in phase and 2 to 1 in amplitude are found from year to year.

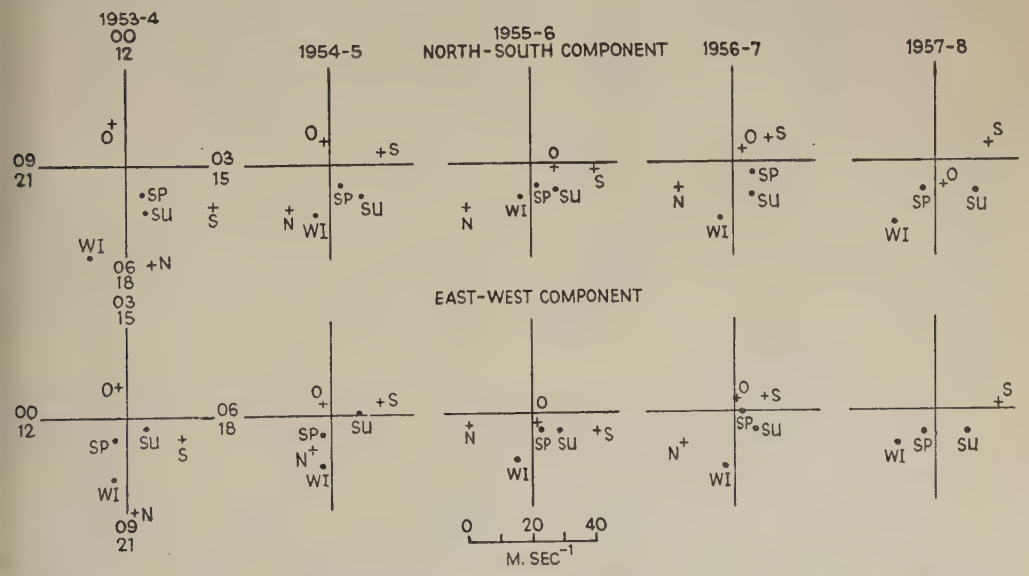

Figure 8. 12-hr harmonic dials for the north-south and east-west components of the semi-diurnal periodic wind variation. Points give amplitude (m sec^{-1}), and local times (GMT) of maximum velocity towards north or east.

● WI – winter; ● SP – spring; ● SU – summer;
+ S – September, + O – October, + N – November.

The months of September, October and November are unusual in that the phase o the 12-hr periodic wind varies markedly. This effect will be discussed later, when th results are considered in greater detail.

The east-west component of the semi-diurnal wind varies in the same way wit season and year as the north-south component, maximum velocity towards east occurrin about three hours later than towards north. Thus to a first order, this wind componen can be represented by a vector of constant amplitude rotating in a clockwise directio twice every day. The observed departures from this simple picture are discussed in th following section.

(ii) *The seasonal variations in phase*

The results of a harmonic analysis for an average day each month, obtained b integration of the five years' observations, are given in Table 2. In the early part of th year maximum velocity towards north occurs at about 0700 hr. The phase then slowl changes at about 10° per month until August, when northward velocity occurs at 0400 hr

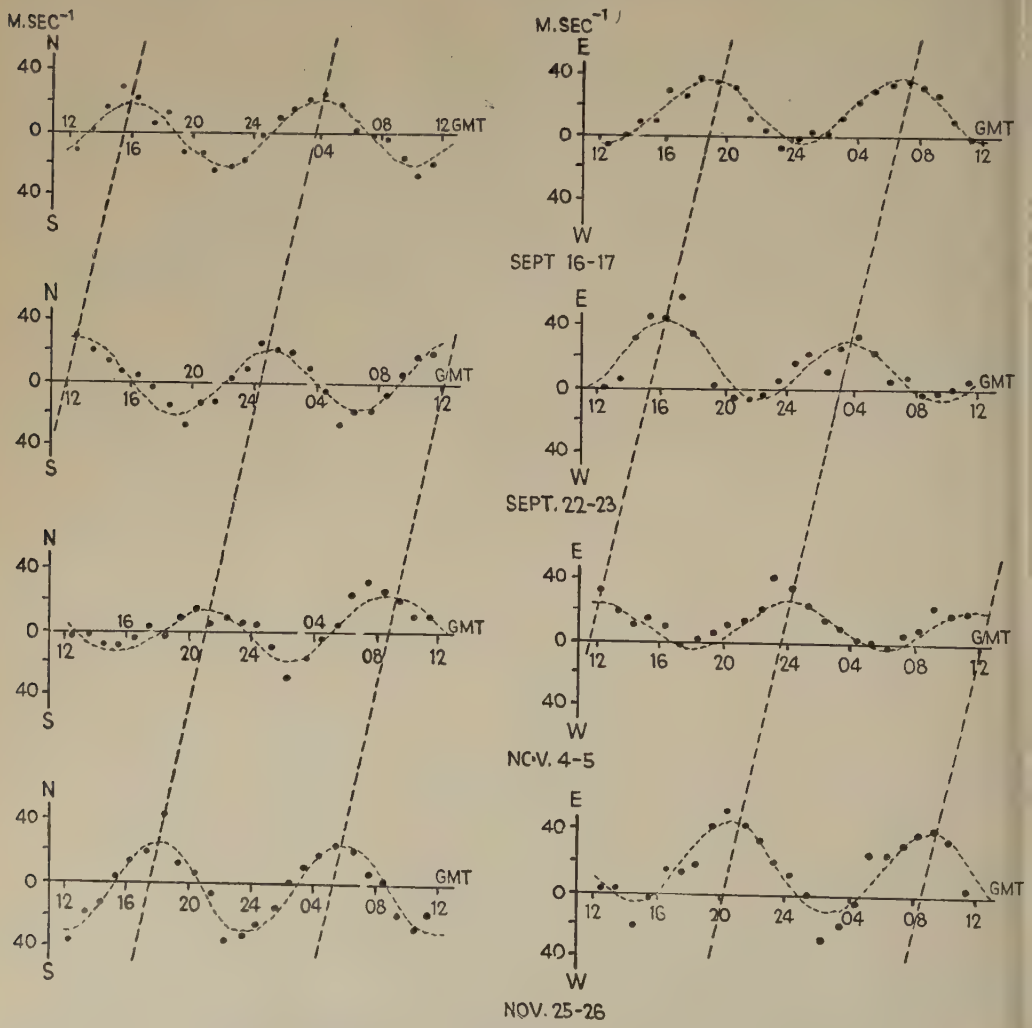

Figure 9. The hour-by-hour variations in the north-south and east-west components of wind velocity for four days in autumn 1954.

The rate-of-change of phase then rapidly increases, maximum northward velocity advancing to 0000 hr by the beginning of October. This is more than a complete phase reversal of 180° compared with January and February. Then, in the course of the next few weeks, the phase retards by 270°, returning in this short time to near the steady winter value. The east-west component is seen to behave in a very similar way. These phase variations are well illustrated by the north-south and east-west wind component plots in Fig. 9, which show the wind measurements on four individual days during autumn, 1954. The progressive changes in the times of maxima of the semi-diurnal wind variation can be seen clearly. It is found that this rapid phase charge does not occur at the same time each year to within a week or so, nor is the rate of change the same. Thus in Table 2, where the months of September, October and November have been sub-divided into shorter intervals, the sub-divisions include those days in each year on which the phases are similar, and are not precisely the same periods of time.

(*iii*) *The phase difference between the north-south and east-west components*

If the semi-diurnal wind variation is caused by a travelling pressure wave of the form

$$\delta p = A(\theta) \sin 2(\omega t + \lambda) \quad . \quad . \quad . \quad (1)$$

where $A(\theta)$ is the latitude variation in amplitude of the oscillation, λ the longitude and ω the angular velocity of the earth, then the wind variation should be represented by a vector of constant amplitude rotating in a clockwise direction in the northern hemisphere. The phase of the rotation should be such that when the pressure is a maximum, the wind blows towards the west. Thus there should be a difference of 90° in phase, or three hours, in the times of maximum north and east velocity. The actual differences in phase for each month are given in Table 2. During ten months of the year the phase difference is very close to three hours, as expected, but large discrepancies occur in May and July. In July, maximum velocity towards north and east occur at almost the same time. This implies that the semi-diurnal wind vector does not rotate during the day, but oscillates sinusoidally in amplitude along a line from roughly north-east to south-west without change in direction. In May, the phase difference is more than three hours, and the variations in the wind vector approximate to an oscillation along a line from north-west to south-east.

TABLE 2. THE 12-HR HARMONIC OF THE DAILY WIND VARIATION

	North-south component		East-west component		$t_2'' - t_1'$
	a_2'	t_2'	a_2''	t_2''	
January	20·0	3·6	19·0	6·6	3·0
February	16·5	4·4	12·5	7·7	3·3
March	5·5	3·0	7·0	6·0	3·0
April	10·5	1·6	6·5	4·4	2·8
May	8·0	2·2	5·0	6·6	4·4
June	10·5	1·8	6·0	4·1	2·3
July	10·0	2·0	6·5	2·6	0·6
August	15·5	1·4	14·0	4·1	2·7
1 September	22·0	0·6	20·0	3·7	3·1
2 September	17·5	10·8	16·5	1·7	2·9
1 October	13·0	8·8	7·0	0·4	3·6
2 October	5·0	2·2	5·5	4·7	2·5
3 October	10·0	5·3	11·0	8·2	2·9
1 November	11·0	5·8	13·5	8·8	3·0
2 November	19·5	4·4	22·5	7·3	2·9
3 November	28·0	2·9	27·0	5·8	2·9
December	25·0	3·0	21·0	5·6	2·6

a_2', a_2'' the amplitudes of the wind components (m sec^{-1});
t_2', t_2'' the phases of the wind components (hr);
the component towards north is given by $V_{S\to N} = a_2' \sin \pi (t - t_2')/6$;
the component towards east is given by $V_{W\to E} = a_2'' \sin \pi (t - t_2'')/6$.

(iv) The amplitudes of the north-south and east-west wind components

The seasonal variations in the amplitudes of the two components of the semi-diurnal wind vector are shown in Fig. 10. In the winter months the amplitudes of the north-south ($V_{S\to N}$) and east-west ($V_{E\to W}$) components are approximately equal, falling from about 30 m sec^{-1} at the end of November to only 6 m sec^{-1} in March. From April to July there is an almost 2 to 1 ratio in amplitudes with $V_{S\to N} = 10$ m sec^{-1} and $V_{W\to E} =$ m sec^{-1}. The amplitudes rise and become equal once more in August and September followed by a sharp dip to only 5 m sec^{-1} during the period of rapid phase variation in October. Thus from August to March $V_{S\to N}$ and $V_{W\to E}$ are very nearly equal in amplitude, and differ in phase by almost exactly 90°. The semi-diurnal wind during this period is, therefore, closely represented by a rotating vector of constant amplitude. In April and June although the phase difference between the north-south and east-west components is approximately three hours, there is a large ratio in amplitude. Thus the locus of the tip of the wind vector is elliptical, and the rotating vector is of variable amplitude with a ratio of 2 to 1 in northerly and easterly directions. As discussed in (*iii*), during May and June the wind vector does not appear to rotate, but oscillates in amplitude along a fixed direction. The results of the harmonic analysis for the north-south and east-west wind components for each month, are given in detail in Table 2.

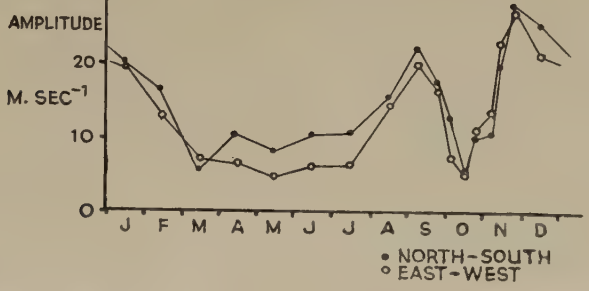

Figure 10. The seasonal variations in amplitude of the north-south and east-west components of the semi-diurnal wind.

(v) Variation with height

The variation with height of the semi-diurnal wind has been discussed previously (Greenhow and Neufeld 1956). It is found that there is a change in both amplitude and phase of this component over the region between 85 and 100 km. The mean amplitude gradient is + 1 m sec^{-1} km^{-1}, rising to + 2 m sec^{-1} km^{-1} in winter and falling to zero during summer. On the average, the amplitude of the oscillation is larger at 100 km than at 85 km by about 15 m sec^{-1}. The mean gradient of phase is approximately + 5° km^{-1} with a maximum value of + 7° km^{-1} occurring in winter and a minimum of 3° km^{-1} in summer. Thus the oscillation at 100 km leads that at 85 km by about 3·5 hr in winter and 1·5 hr in summer. The observations described in this paper refer to a mean height of 92 km, and the results may be extrapolated to other heights in the meteor region using these measured gradients of amplitude and phase.

(c) *The 24-hr periodic wind*

(i) Seasonal variations

The diurnal wind component in the 90 km height region appears to be very variable in amplitude. Usually it is smaller than the more regular semi-diurnal wind vector, although sometimes, because of this variability, it may exceed the 12-hr wind in amplitude.

The phase of this component also changes considerably from day to day, and deviations of $\pm 90°$ from the mean seasonal values are commonly observed.

Because of the random variations in the amplitude and phase of the 24-hr periodic wind, the observations for all the days in the five years from September 1953 to August 1958 have been averaged for each season, in order to resolve any systematic variations. The results of the harmonic analysis are given in Table 3.

TABLE 3. THE 24-HR HARMONIC OF THE DAILY WIND VARIATION

	North-south component ($V_{S\to N}$)	East-west component ($V_{W\to E}$)
Spring	$7{\cdot}8 \sin \pi (t - 6{\cdot}5)/12$	$2{\cdot}3 \sin \pi (t - 10{\cdot}6)/12$
Summer	$9{\cdot}2 \sin \pi (t - 5{\cdot}8)/12$	$3{\cdot}7 \sin \pi (t - 15{\cdot}9)/12$
Autumn	$4{\cdot}2 \sin \pi (t - 6{\cdot}7)/12$	$3{\cdot}4 \sin \pi (t - 9{\cdot}7)/12$
Winter	$3{\cdot}6 \sin \pi (t + 3{\cdot}7)/12$	$3{\cdot}9 \sin \pi (t + 0{\cdot}8)/12$

Spring – March, April, May. Autumn – September, October, November.
Summer – June, July, August. Winter – December, January, February.
Amplitudes (m sec^{-1}); time is measured from 0000 hr local time (GMT).

In spring and summer the amplitude of the north-south component is approximately 8 m sec^{-1}, maximum velocity towards north occurring at 1200 GMT. The amplitude falls to only 4 m sec^{-1} in autumn although the phase remains the same. During the winter months, however, a complete phase reversal takes place, maximum velocity towards north occurring at about 0200 hr with an amplitude of 4 m sec^{-1}. The amplitude of the east-west component during spring, summer and autumn is less than half that of the north-south component, whilst winter amplitudes are equal. On the average, the time of maximum velocity towards east occurs approximately 6 hr after maximum northward velocity (a 90° phase difference), showing that the wind vector rotates in a clockwise sense with a period of 24 hr. The actual phase differences for the individual seasons differ appreciably from 6 hr, however, so that the rate of rotation of the wind vector is not constant. As the amplitudes of the north-south and east-west components are not equal, the result is that the tip of the wind vector traces out a very elliptical path throughout the day.

(*ii*) *Comparison of four years' observations*

In order to investigate year-to-year variations of the diurnal wind component, the observations for all days in each of the years from September 1953 to August 1957 have been averaged together, irrespective of season. About 100 days have been used. The mean yearly values of amplitude and phase of the north-south component are shown in Fig. 11. There is good agreement between the four years' observations showing that on the average, the 24-hr periodic wind follows the same behaviour from year to year. The amplitude varies by only $\pm$ 20 per cent, and the phase by $\pm 20°$ from the overall mean values. Because of the seasonal variations in amplitude and phase of the east-west component, the resultant amplitude is nearly zero when averaged over a year. For this reason, yearly means of the east-west component do not have much significance, and have not been compared.

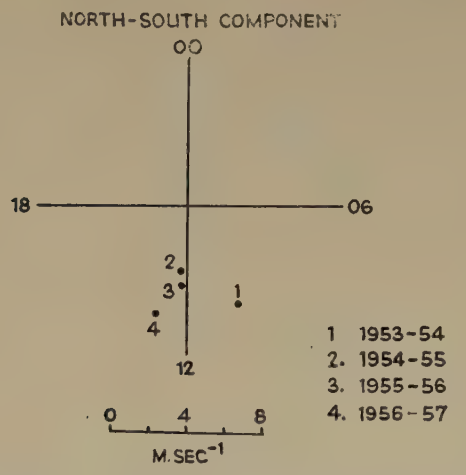

Figure 11. Comparison of the north-south component of the diurnal wind variation in four successive years

(d) *The 8-hr periodic wind*

(i) *Resolution of the 8-hr component*

At the surface of the earth there exists a significant 8-hr periodic variation in the daily pressure variation. A search has been made for a corresponding 8-hr. periodic wind system in the upper atmosphere. Yearly averages of the north-south and east-west wind components have been taken, and the harmonic analysis extended to include the third harmonic. An example of the results is shown in Fig. 12, to demonstrate the significance of the 8-hr wind component which is resolved in this way.

Fig. 12 (a) shows the average north-south component of the measured wind for the year 1955. The broken line represents the sum of the first three terms of the Fourier series.

$$V_{S\to N} = -3{\cdot}2 + 2{\cdot}8 \sin \pi (t - 6{\cdot}2)/12 + 8{\cdot}1 \sin \pi (t - 2{\cdot}6)/6$$

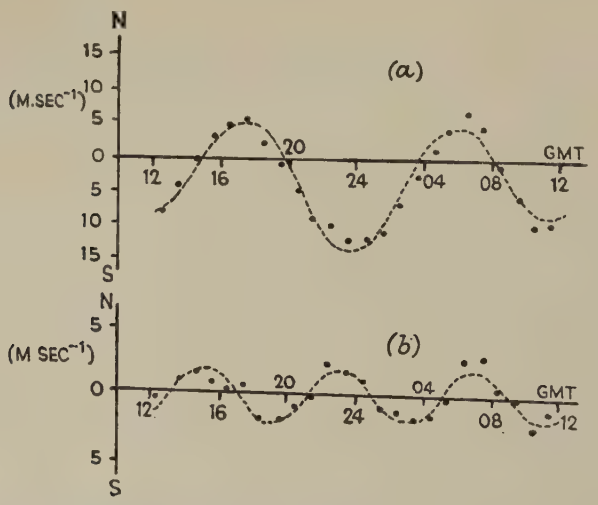

Figure 12. The mean north-south wind component for the year September 1954 to August 1955, showing the presence of a significant 8-hr periodic variation.

where the constant term gives the value of the prevailing wind, 3·2 m sec^{-1} towards south, and the second term the diurnal wind component with an amplitude of 2·8 m sec^{-1} and phase (maximum velocity towards north) of 12·2 hr. The 12-hr component has an amplitude of 8·1 m sec^{-1} averaged over a year, maximum northward velocity occurring at 5·6 hr and 17·6 hr.

The points for the wind velocity in each hour show significant departures from the broken curve in Fig. 12 (a). These deviations are plotted in Fig. 12 (b), where they are seen to form a regular and quite significant 8-hr oscillation of the wind given by

$$V_{S\to N} = 2{\cdot}1 \sin \pi (t - 5{\cdot}1)/4$$

The amplitude of this component is 2·1 $\pm$ 0·2 m sec^{-1}, maximum velocity towards north occurring at 7·1, 15·1 and 23·1 $\pm$ 0·1 hours. These deviations from the smooth curve in Fig. 12 (b) are just those expected from random errors in the individually hourly wind measurements ($\sim$ 0·6 m sec^{-1}). No wind components with periods less than 8 hr exist with amplitudes greater than approximately this value.

(ii) Seasonal and yearly variations

The mean values of the north-south and east-west wind components for each of the three years from 1954 to 1956, have been analysed for 8-hr harmonics in the way described in the previous paragraph. The amplitudes and phases of the wind components which have been resolved are given in Table 4. In the north-south direction the mean amplitude of the wind vector is about 2·5 m sec^{-1}, maximum velocity towards north occurring at 0730 hr. The values of the amplitude in each of the three years are similar, and the phases agree within $\pm$ 45°. The mean amplitude in the east-west direction is 3 m sec^{-1}, with maximum eastward velocity at 0900 hr. Thus the amplitudes of the north-south and east-west components are similar, and the phase difference is approximately 90°, showing that the wind component can be represented by a clockwise rotating vector of constant amplitude.

TABLE 4. THE 8-HR HARMONIC OF THE DAILY WIND VARIATION.

	North-south component ($V_{S\to N}$)	East-west component ($V_{W\to E}$)
1954	2·4 sin $\pi (t - 6{\cdot}7)/4$	1·7 sin $\pi (t - 7{\cdot}0)/4$
1955	2·6 sin $\pi (t - 5{\cdot}0)/4$	4·3 sin $\pi (t - 6{\cdot}7)/4$
1956	3·6 sin $\pi (t - 5{\cdot}2)/4$	3·5 sin $\pi (t - 7{\cdot}2)/4$
Mean	2·5 sin $\pi (t - 5{\cdot}5)/4$	3·1 sin $\pi (t - 7{\cdot}0)/4$

Amplitudes (m sec^{-1}); time is measured from 0000 hr local time (GMT).

The seasonal variation of the third harmonic has been investigated by averaging together the three years' observations in four three-monthly groups. The results for the east-west component are shown in Fig. 13. The points for the four seasons are clustered in one quadrant of the harmonic dial, and the amplitudes and phases do not differ significantly from the mean value given in Table 4. Any seasonal variation is therefore less than $\pm$ 20 per cent in amplitude and $\pm$ 20° about the mean value.

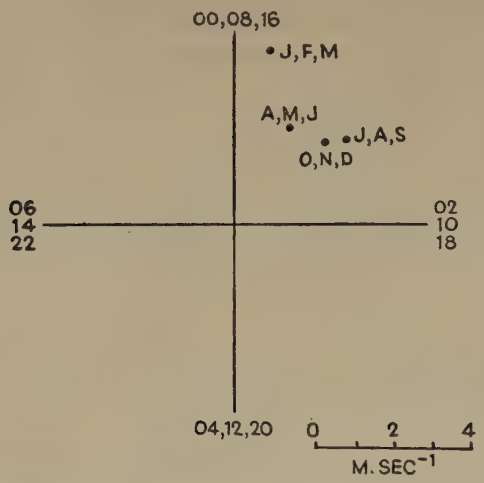

Figure 13. Seasonal variations in amplitude and phase of the 8-hr wind component.

(e) *The lunar semi-diurnal wind component*

A search for the lunar semi-diurnal wind system, which will be produced by the tidal action of the moon, has been made by averaging together in lunar time all the observations obtained during the five years from 1953 to 1958. Zero of time has been taken at lower culmination of the moon. The residual north-south and east-west wind components obtained in this way are shown in Fig. 14. Significant second harmonic terms with amplitudes of approximately 2 m sec^{-1} remain, as shown by the fits of the broken curves to the experimental points. It is, however, necessary to distinguish between a genuine lunar semi-diurnal wind component, and a residual solar component due to the random integration of the 100 or so days measurements. An estimate of the uncertainty in the final result has been obtained by considering each of the four years separately. The 12-hr components in lunar time are shown by the harmonic dial in Fig. 15. The east-west and north-south components are displayed on the same harmonic dial by displacing the east-west phases through 3 hr. The points for each year are considerably scattered about the mean value, and the error circle deduced from this scatter, is shown in the figure. It can be seen that the error in amplitude of the apparent lunar semi-diurnal wind is equal to its amplitude. Thus the measurement is of doubtful significance, and serves to set an upper limit of 2 m sec^{-1} to the amplitude of this wind component.

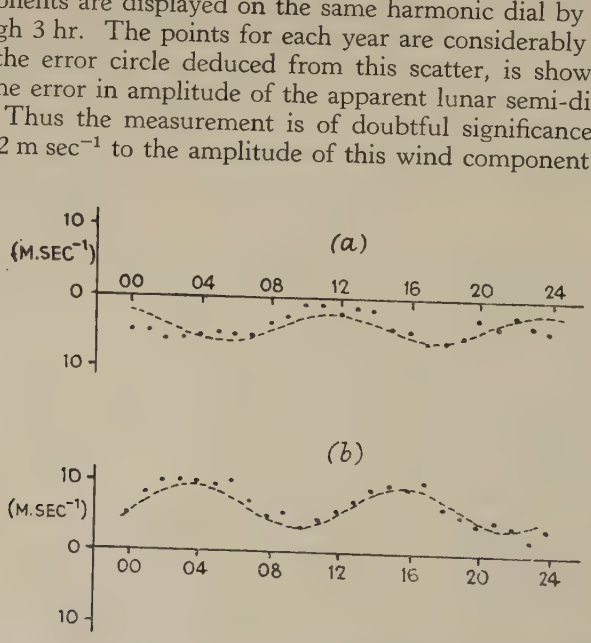

Figure 14. North-south (a) and east-west (b) components of the wind, averaged in lunar time.

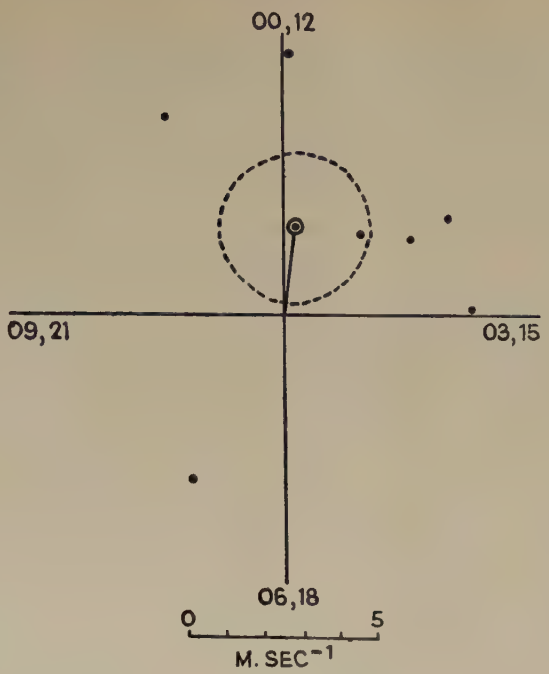

Figure 15. The lunar 12-hr harmonic dial (⊙ – mean value).

4. Discussion

(*a*) *The prevailing wind*

The pattern of the prevailing wind at 80-100 km, consisting of large velocities towards south-east in summer, eastward winds in winter and small westward velocities in spring and autumn, appears to be quite stable. There is good similarity between the seasonal variations for each of the five years of measurement. It is possible to interpret the mean wind in terms of horizontal and vertical temperature gradients, and the zonal east-west winds are explicable by considering the effect of a temperature difference between the polar and equatorial regions. The temperature gradient necessary is approximately 1°K per 1,000 km, taking known values of the vertical wind gradient (Greenhow and Neufeld 1956) and of the temperature gradient (Rocket Panel 1952). The results require that during summer and winter the temperature is lower at the poles than at the equator with the reverse at the equinoxes. The north-south wind is more difficult to explain, and presumably there is a return air stream at a different height or in different longitudes.

The only observations directly comparable with those described in this paper have been made by Elford and Robertson (1953) and Elford (1959) at Adelaide. These workers also used a meteor-echo technique, and the measurements refer to a latitude 35°S. Their results showing the height variation of the east-west component for four months at quarter-yearly intervals, are shown for comparison in Fig. 6 (b). A number of similarities between the two series of observations can be seen. The northern hemisphere spring results (represented by April and May) are similar to those in the southern hemisphere spring (September), with westward winds and small height gradients. The autumn results represented by September and March in the two hemispheres are also similar, with eastward winds and small height variations. Eastward winds with similar positive and negative gradients are observed at other times of the year at both stations, but the Adelaide results are not sufficient to show whether there is an annual or a bi-annual variation of the wind in the southern hemisphere.

The curve for December-January in the northern hemisphere has no counterpa in the southern hemisphere. This may be due to a real difference in the winds, but the other hand, may be due only to the absence of observations from Adelaide for th months of July and August.

A major difference between the winds measured at Jodrell Bank and Adelaide that the velocities observed in the southern hemisphere are about three times as high the northern hemisphere measurements. The wind gradients also differ in the same rati

(*b*) *The semi-diurnal wind*

(*i*) *The pressure oscillation*

At the surface of the earth a semi-diurnal pressure variation is detectable with a amplitude δp approximately one-thousandth of the steady atmospheric pressure Maximum pressure occurs at 1000 and 2200 hr (local time), although there is a small seasona variation in phase of about 30° (Chapman 1939) and see Fig. 16 (b). The theoretic amplitude of the wind due to this pressure oscillation is about 0·4 m sec^{-1}. This sma 12-hr wind component is obscured by the much larger irregular winds which depend o local surface features, although it has been detected at a height of about 15 km (Johnso 1955).

The semi-diurnal pressure oscillation is caused by thermal and tidal effects of the sur At greater heights it is suggested that the amplitude of the oscillation is increased man times by resonance (Taylor 1936; Pekeris 1937). Wilkes (1949) has used an approximat temperature profile of the atmosphere, on which the oscillation depends critically, t predict the variation in amplitude and phase of the semi-diurnal pressure oscillation with height. The amplitude and phase of the pressure oscillation at 92 km may be deduce from the known wind structure, using the equations of motion for a particle in a rotatin atmosphere. The seasonal variations in amplitude and phase of $\delta p/p$ for a height of 92 km assuming an idealized pressure wave of the form given by Eq. (1), is shown in Fig. 16 (a). During the winter months the phase differs from the surface oscillation by approximately 180°, as predicted by Wilkes (Greenhow and Neufeld 1956), but very large deviation occur at other times of the year. During the period of rapid phase change in autumn the oscillation at 92 km is, for a time, in phase with that at the ground. It is difficult however, to postulate mechanisms other than resonance which satisfactorily explain the large 12-hr wind component in the upper atmosphere. Any direct absorption of sola energy in the meteor region itself would not be expected to show the erratic changes in

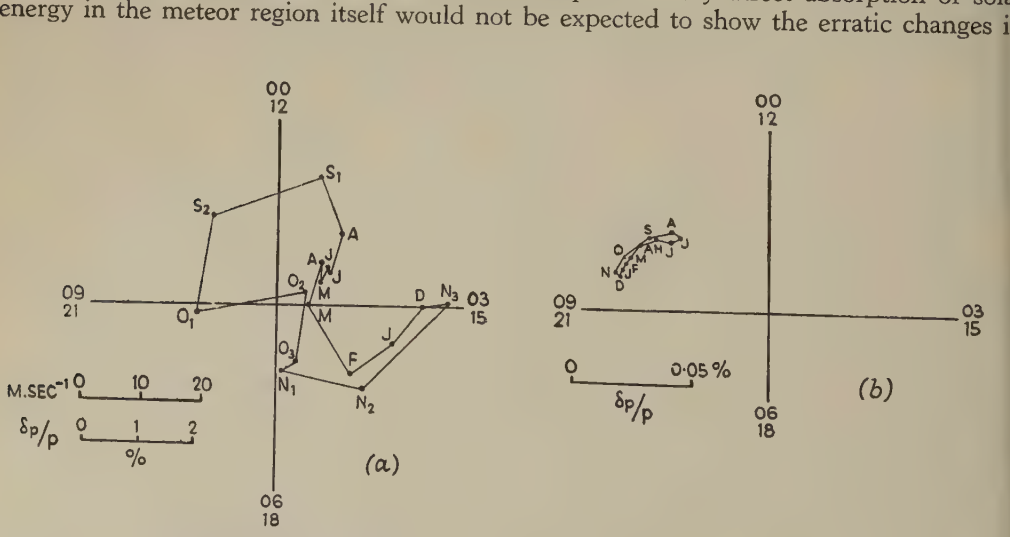

Figure 16. (a) The pressure oscillations $\delta p/p$ at 92 km (meteor wind measurements) and (b) at the surface of the earth (Chapman 1939).

hase and amplitude which occur and in any case would be largely diurnal in character. n occasions there are random changes in phase of about 30°, and a factor two in mplitude, from one day to another, with similar changes from one year to another. hese variations however, are not unreasonable if the atmospheric resonance is sufficiently harp to give a magnification of 40 or so, shown by comparison of Fig. 16 (a) and (b). he predicted period of resonance depends critically upon the form of the temperature istribution in the upper atmosphere, and regular seasonal variations together with a andom day-to-day and year-by-year variation in temperature could explain the observed ehaviour of the semi-diurnal wind component.

It is of interest to note that the small seasonal variation in the phase of $\delta p/p$ at the round shows the same trend as the variation at 90-100 km. There is a gradual variation f 30° from November through to August, and then a sudden return during the autumn nonths. Thus to a lesser extent, the rapid phase changes during autumn in the upper tmosphere are present at the surface of the earth.

ii) Comparison with other observations

Drift measurements in the lower E-region using the Mitra (1949) three-station ading technique have been made at a number of places in the northern hemisphere, nd the 12-hr harmonic has been resolved (Phillips 1952; Chapman 1953; Briggs and pencer 1954; Harang and Peterson 1957). These measurements, which refer to a height ather greater than the meteor observations (105-110 km), show the presence of a semi-diurnal wind component of about 30 m sec^{-1}. The wind vector is directed towards north at about 3·5 hr (local time), which is about two hours ahead of the winds in the meteor region. If the known values of the height variation of the amplitude and phase from the meteor observations are taken into account, the measurements by the two different methods become consistent with one another. Recently Jones (1958) at Cambridge, working on two slightly different frequencies has obtained a height variation for the 12-hr component. His results show a phase gradient of about 7° km^{-1}, in good agreement with the meteor observations. Jones also performed a more detailed analysis of the earlier Cambridge results during autumn, in order to search for the rapid phase changes observed in the meteor region. A large phase change over this period was detected, the phase becoming earlier during the period. Similar phase changes were obtained on re-analysing Phillips's results for 1949, 1950, 1951. The unusual behaviour of the 12-hr wind component during autumn is thus present over a wide height range.

The measurements made using the meteor-echo technique at Adelaide show that at lat. 35°S the 12-hr component is very irregular in amplitude and phase, and smaller than the corresponding 24-hr component. At times the wind approximates to an anti-clockwise rotating vector, as expected in the southern hemisphere, but it is extremely unstable. During some months there is a phase change with height, with the higher levels leading the lower ones.

At the latitude of Adelaide, therefore, it appears that the main periodic component is represented by a large 24-hr harmonic. This is the reverse of the position at Manchester, where the 24-hr harmonic is generally much smaller and more irregular than the semi-diurnal wind.

(*c*) *The 24-hr component*

At the surface of the earth the amplitude of the diurnal pressure wave is of the order 0·3 to 0·5 that of the semi-diurnal oscillation, although in general the 24-hr wave is irregular compared with the 12-hr pressure system. The amplitudes of the 12-hr and 24-hr wind systems should therefore be approximately equal. At a height of 92 km and lat. 53°N, the amplitudes of these two periodic winds are of the same order in spring and summer, although as at ground level, the phase of the 24-hr component is very irregular. In winter, however, the semi-diurnal wind is at least five times bigger than the diurnal wind. This behaviour is not found in the wind measurements made at Adelaide, lat. 35°S, which are

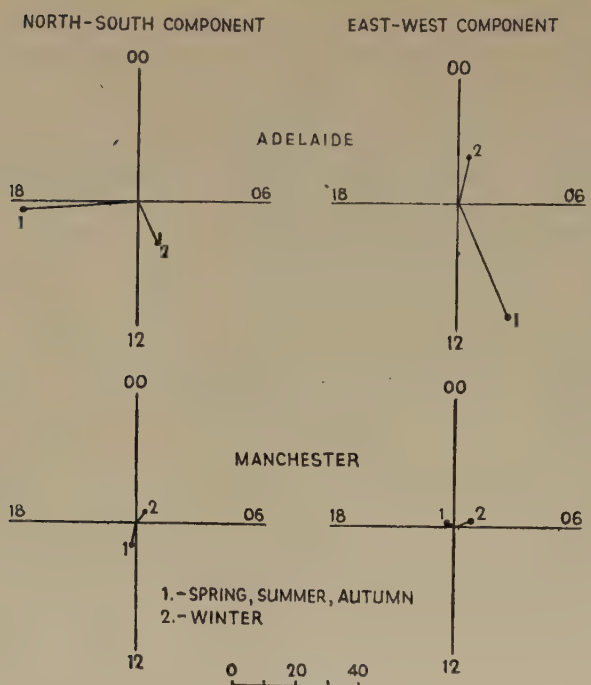

Figure 17. Comparison of the diurnal wind component at Adelaide and Manchester.

the only observations with which the present results may be compared. At Adelaide th 24-hr wind appears to be the major component, and the semi-diurnal wind is somewha irregular. The two series of observations are compared in Fig. 17. The harmonic dial showing the Adelaide measurements are based on results given by Huxley (1957). Southern hemisphere winter is taken as the month of June, summer as December, spring as Septembe and October, and Autumn as March. It can be seen the amplitude of the north-south component at Adelaide (~ 40 m sec^{-1}) is approximately five times as big as that a Manchester (~ 8 m sec^{-1}). A diurnal wind of 40 m sec^{-1} would be produced by a 10 per cent pressure oscillation. A pressure variation of this order has been measured at Manchester (Greenhow and Hall 1960) although the wind speed is not correspondingly high. However, calculations of the wind speed are made assuming a simple travelling pressure wave with a maximum at the equator, falling off in some assumed manner to zero at the poles. Such a pressure wave would give rise to a wind vector of almost constant amplitude, rotating with uniform angular velocity clockwise in the northern hemisphere and anti-clockwise in the southern hemisphere. This idealized behaviour is not followed by the diurnal wind vector. Although the directions of rotation are correct as shown by the time displacements of maximum northward and eastward velocity in Fig. 17 the rates of rotation, and amplitudes, are very variable. Fig. 17 also shows that the times of maximum northward velocity are displayed by 6 hr ($\pi/2$ in phase) in the two hemispheres, whereas with a simple travelling wave the phase difference would be zero or π. The two series of measurements are similar in that the phase of the diurnal wind in winter is opposite to that in spring, summer and autumn, for both the north-south and east-west components.

(*d*) *The 8-hr component*

The results given in Table 4 show that the north-south and east-west components of the third harmonic of the wind system are approximately equal, and displaced by 90° in phase. The 8-hr wind component is therefore quite well represented by a vector with a

nstant amplitude of 2 m sec^{-1}, rotating in a clockwise direction, such as would be pected from a travelling pressure wave. The phase of the oscillation appears to be ite stable. The amplitude of the 8-hr component is about 0·2 times the amplitude of e semi-diurnal wind averaged over a year. This is of the order expected from the ratio amplitudes of the corresponding pressure oscillations at the surface of the earth, which ries from 0·1 to 0·2.

(e) *The lunar semi-diurnal wind*

Several attempts have been made to evaluate the lunar 12-hr periodic wind component, iefly using the Mitra three-station technique at heights of approximately 105 km hillips 1952; Chapman 1953; Briggs and Spencer 1954; Harang and Peterson 1957). he measurements are all of doubtful significance, and essentially set an upper limit about 10 m sec^{-1} for the amplitude of the lunar component. The present measurements duce this limit to 2 m sec^{-1}. From the amplitude of the lunar pressure oscillation redicted by Wilkes (1949), a wind component of about 0·5 m sec^{-1} would be expected. his is much smaller than the amplitude which can be resolved with the present data.

Acknowledgments

The work described in this paper has been carried out at the Jodrell Bank Experimental Station, University of Manchester. The authors wish to thank Professor A. C. B. Lovell, F.R.S., Director of the station, for his interest in the investigation.

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Total ozone and perturbations in the middle stratosphere

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(Manuscript received 21 December 1960)

SUMMARY

Total ozone amounts are shown to vary with the motion systems of the stratospheric polar vortex. Tw cold outbreaks, one over Europe in 1959 and the other over North America in 1960, produced similar pro nounced minima in the total ozone records. The results support the hypothesis that major anomalies in t horizontal and seasonal distributions of total ozone are related to dynamical activity in the polar-night vort and the stratospheric warm belt.

1. INTRODUCTION

For many years it has been well known that the total ozone in a vertical column o unit cross-section is subject to large short-period variations. Since ozone concentration are almost certainly in photochemical equilibrium above about 35 km (Craig 1950), an are small in the troposphere, these short-period variations must be due to processes i the lower and middle stratosphere. Reed (1950) suggested a dynamical mechanism based on the idea that subsidence in the lower stratosphere should increase total ozone A recent formulation by Godson (1961) shows that Reed's mechanism will usually hav the required sign at levels below the maximum of ozone mixing ratio (35-45 km). Fig. shows the normal profiles.

The normal horizontal temperature gradients of the lower stratosphere in temperat and polar latitudes are small, and the strong positive vertical gradients of potential tem perature play the dominant role in the thermal structure. Warm and cold regions ar created adiabatically by vertical circulations induced by tropospheric cyclogenesis and anticyclogenesis, subsidence bringing down potentially warmer air and ascent doing th reverse. Since the tropospheric disturbances decrease in amplitude above the tropopause hydrostatic considerations require the ridges to be cold and the troughs warm. The stratospheric warm and cold regions move and change in intensity in such a fashion as to remain approximately in phase with the generating lower-level wave regime. In the active regions of the lower stratosphere the wind speeds are much greater than the phase velocities of the disturbances; hence the air subsides as it flows from the cold ridge to the warm trough and ascends as it moves from the trough to the ridge. These thermal anomalies are most strikingly demonstrated on 200-mb charts.

The quasi-conservative ozone mixing ratio increases with height above the tropopause, as does potential temperature. Hence it is to be expected that ozone anomalies will arise, move and change in intensity relative to the tropospheric perturbations in the same manner as 200-mb temperature anomalies. Troughs are likely to have high amounts of ozone and ridges lower amounts. These ozone-tropospheric perturbation effects have been well substantiated by Normand (1953).

The upper boundary for the effect of these induced vertical circulations, at least for the summer half of the year, varies very little in time or space, being at about 50 mb (Hare 1960a). However, the lower boundary of the ozone-rich layer varies considerably in height, and it is difficult to demonstrate the Reed-Normand effect quantitatively; even at its best it appears to explain only about 25-35 per cent of the daily variance of total ozone, and in winter it sometimes fails entirely.

Inspection of ozone time-series in winter often reveals the existence of large amplitude variations of rather longer period than the type invoked by Reed (the latter changes are due to the eastward movement of wave perturbations in the tropospheric or *Ferrel* westerlies (Hare 1960b)). Godson (1960) and Allington, Boville and Hare (1960) showed that these

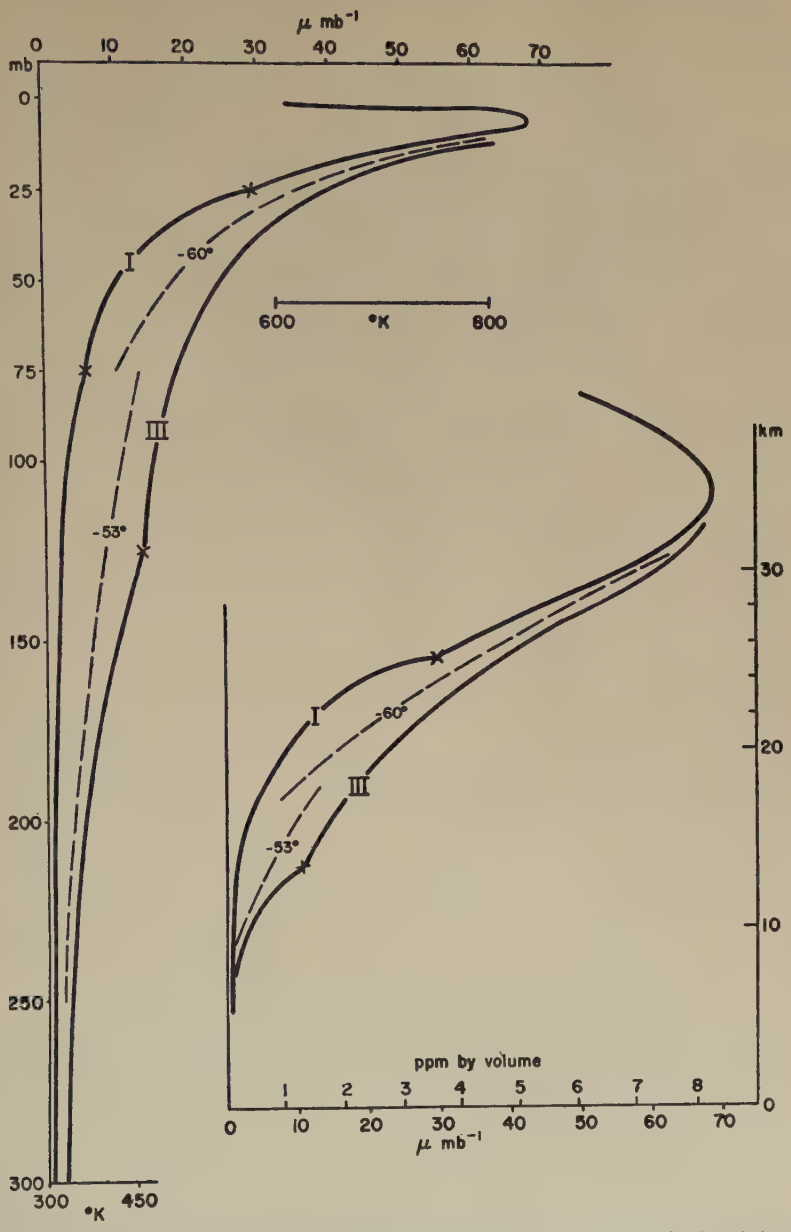


Figure 1. The vertical distribution of ozone, microns per millibar layer, for standard *umkehr* curves I and III, together with profiles of potential temperature in an isothermal stratosphere at — 53C and — 60C.

larger variations could be referred to travelling waves in the polar-night westerlies of the middle stratosphere. Low total ozone values occur in phase with the troughs of this upper vortex, and high values in phase with ridges, the ozone isopleths appearing to move at the phase velocity of the associated thermal waves. The correlations obtained by the second group of writers (at Edmonton and Moosonee) explained about 50 per cent of the ozone variance when the polar-night waves were present. These results suggest that very large anomalies shown by winter ozone time-series in mid-latitudes may be due to the southward movement of polar-night troughs, and their cutting-off as cold vortices. The work now presented is designed to test this hypothesis, and to cast some light on the general control of ozone concentrations.

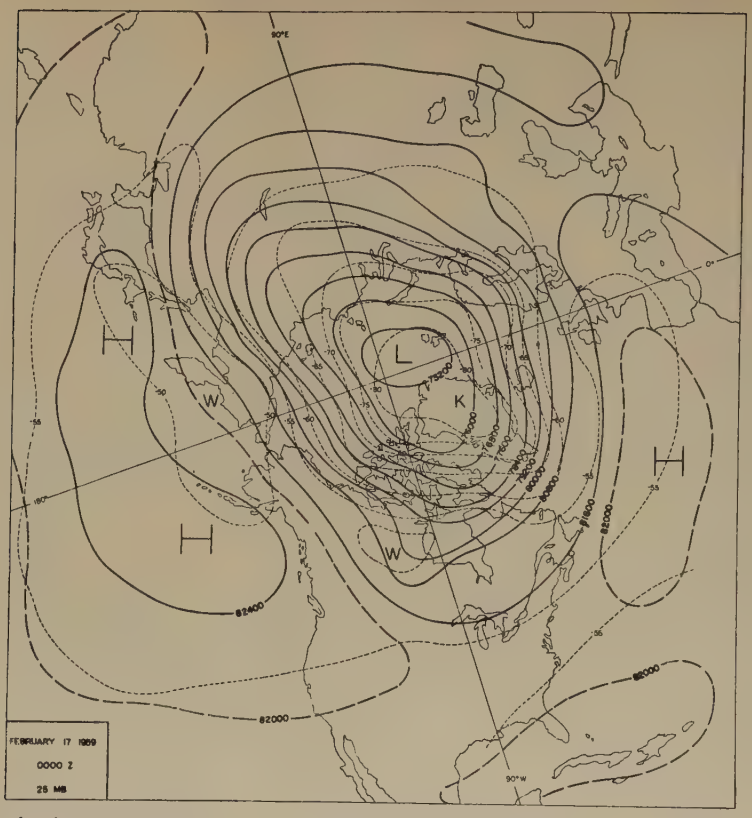


Figure 2. 25-mb circumpolar chart for 17 February 1959, at 0000 hr GMT. Contours of geopotential height (in gpft.) solid lines; isotherms (deg C) dashed. Note troughing of isotherms over Greenland.

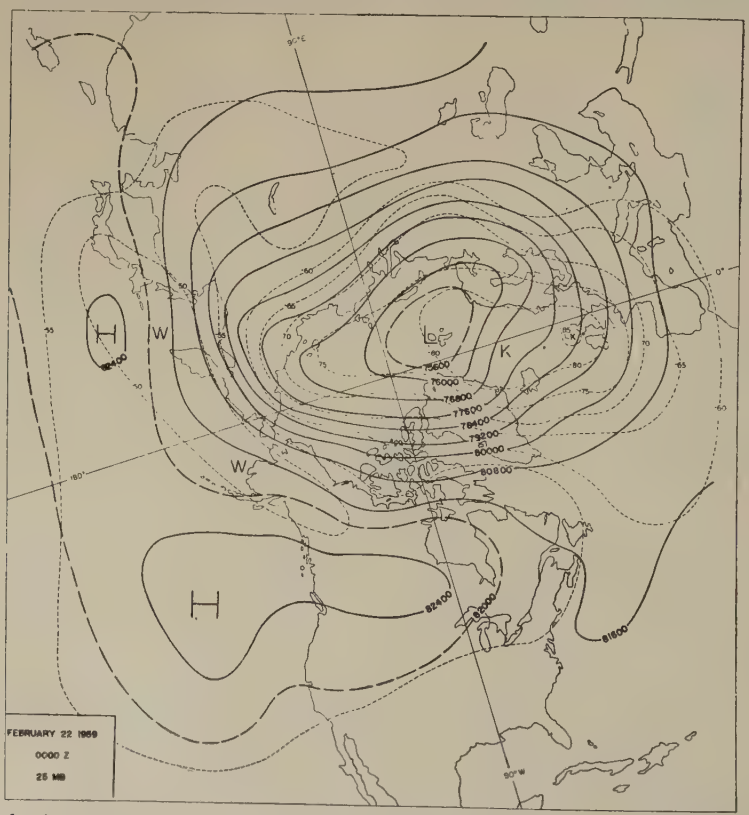


Figure 3. 25-mb circumpolar chart for 22 February 1959, at 0000 hr GMT, uniform with Fig. 2. Note trough and closed — 85C isotherm over British Isles.

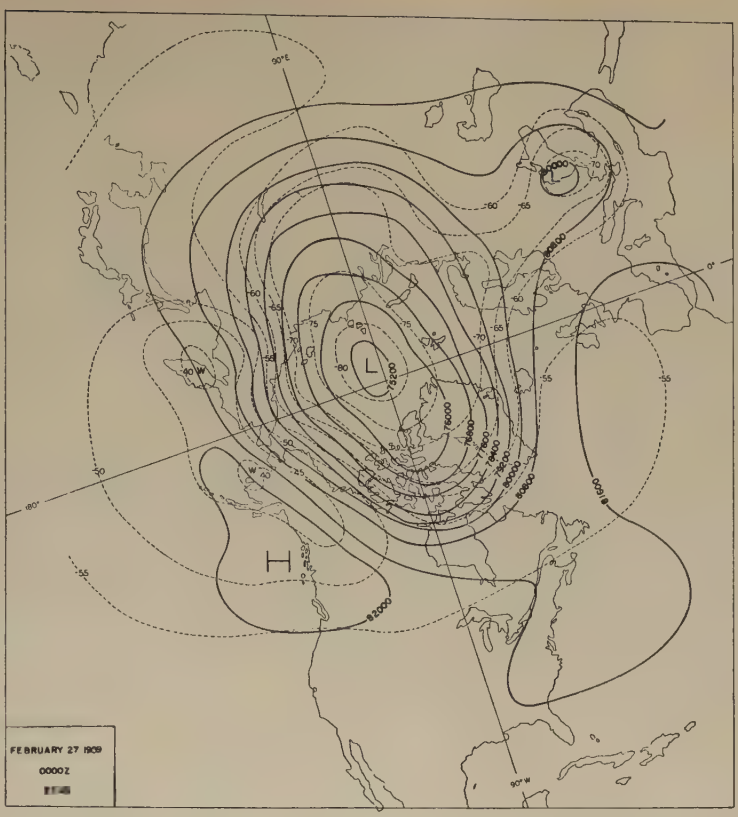


Figure 4. 25-mb circumpolar chart for 27 February 1959, at 0000 hr GMT, uniform with Fig. 2. The trough over Britain in Fig. 3 has become a closed cold low over the Black Sea, having moved south-east across Germany.

2. Data employed

Daily total ozone values were obtained for the period July 1958 to January 1960 for a selection of North American and European Dobson spectrophotometer stations; specific use is made of the daily records from Lerwick, Oxford, and Rome in Europe and Moosonee in Canada. The degree of accuracy of these records is well known, and no further justification will be presented here. The unit employed for total ozone is the micron, μ, per unit area at S.T.P. (10^{-4} atm cm, Vigroux scale). Use is made of the daily chart series for stratospheric levels maintained by the Arctic Meteorology Research Group, McGill University (200 mb, 100 mb and 25 mb; 30 mb after 1 January 1960). The 25-mb maps for the 1958-59 winter were extended, every fifth day, to a circumpolar coverage, and the history of each perturbation was traced by means of isochrone maps. In addition, time cross-sections to the top of the rawinsonde record were maintained for a selection of stations in North America and Europe. Examples of these maps and cross-sections form Figures 2-4.

3. *Case studies*: Effect of stratospheric cold lows

The hypothesis to be investigated was that the large, unexplained ozone variations in mid-latitudes might be due to the southward movement of cut-off cold lows from the polar night westerlies. During the 1958-59 and 1959-60 winters, two such cases were isolated and investigated; no other was observed. The first occurred over the North

Atlantic and Europe during February 1959, and will be referred to as *Case A*. Th second occurred over North America during January 1960, and is referred to as *Case E* Both systems produced very spectacular ozone minima along their paths.

Case A

On 17 February 1959, (Fig. 2) the polar-night vortex was centred north of Greenland A baroclinic trough over Greenland began to deepen, and then moved south-eastward By 22 February (Fig. 3) the thermal centre was closed, and situated over Northern Ireland, moving east-southeastward at about six degrees of longitude per day; on 22 February (Fig. 4) it was over the Black Sea. It produced 30-mb temperatures of — 88C at Stornoway and — 89C at Leuchars (Scotland). Even over the Black Sea, 30-mb temperatures fell below — 70C. The time cross-section for Bitburg, Germany, for February 1959 (Fig. 5) shows that the system's passage caused a profound cooling through most of the stratosphere, followed by a warming to normal values.

Very large total ozone changes occurred as the system passed. Fig. 6 gives time sections for total ozone and 30-mb temperature at Lerwick, Oxford and Rome, the temperature record on the Oxford section being for Camborne (Cornwall). At all three stations the records show (i) short-period fluctuations of amplitude up to 700 microns uncorrelated with 30-mb temperature; (ii) a very pronounced minimum associated with the passage of the stratospheric cold core on 22 February at Lerwick, 23 February at Oxford, and 25 February at Rome; (iii) a sharp recovery to more normal values in the rear of the core, followed at Lerwick only by a rise to very high values (5,350 microns on 6 March, associated with the final stratospheric warming of the 1958-59 winter).

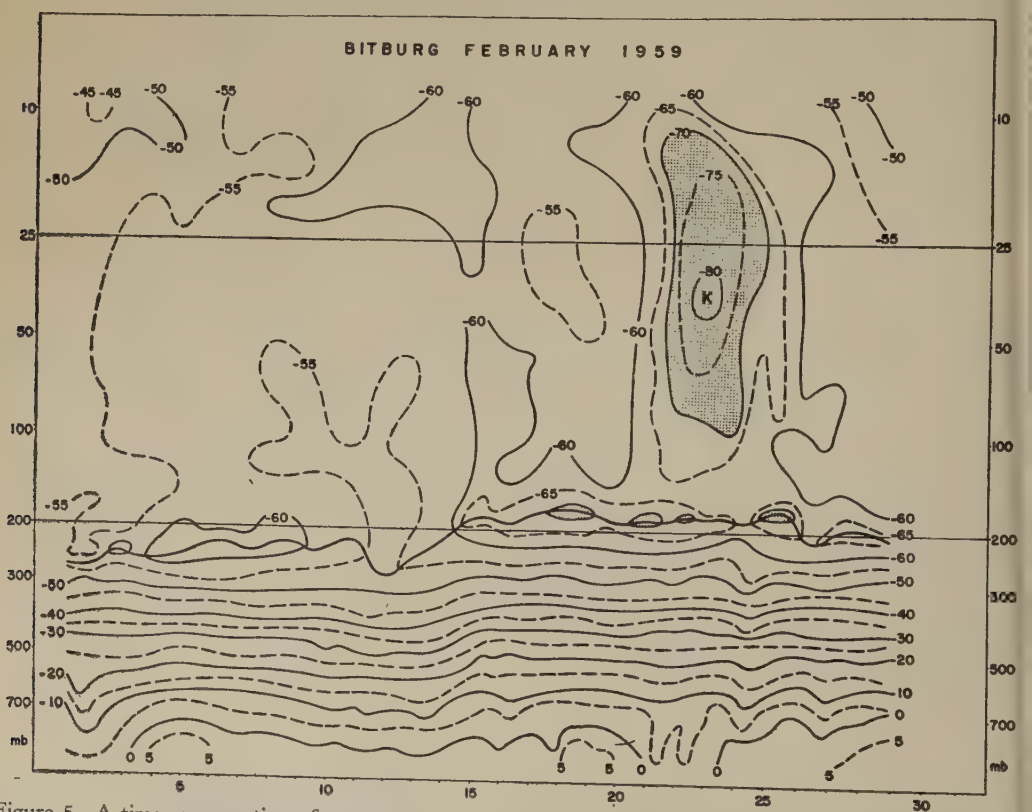


Figure 5. A time cross-section of temperature (deg C) over Bitburg, West Germany, for February 1959. The coordinates are pressure (mb) and time (date). For most of the month temperatures in the stratosphere lay in the — 55C to — 60C range, but a profound cooling (to below — 80C at 30 mb) occurred during the passage of the cold trough shown in Fig. 4. See Fig. 6 for the effects of this disturbance on ozone amounts.

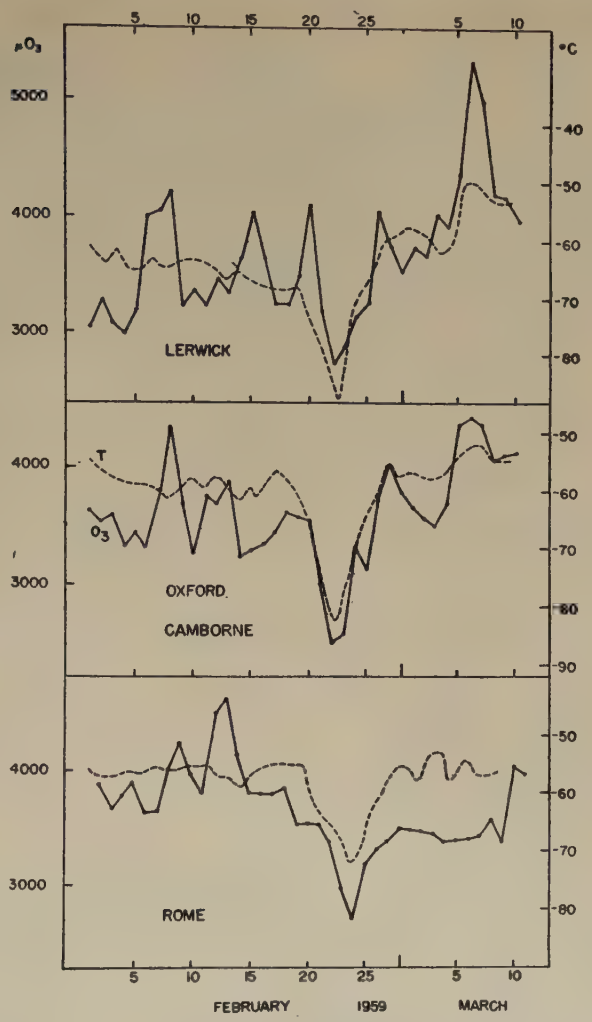


Figure 6. Time sections of 30-mb temperature (deg C, dashed lines) and total ozone (microns, solid lines) at Lerwick, Oxford (Camborne temperatures) and Rome, 1 February to 10 March 1959. The curves show the sharp minima in total ozone and temperature associated with the passage of the perturbation shown in Figs. 2-5.

The variations listed under (i) above can be ascribed to travelling waves in the tropospheric (Ferrel) westerlies, i.e., to the Reed-Normand effect. Those under (ii), however, are of larger amplitude, and are connected with the dynamical effects of the stratospheric cold core. At all three stations the total ozone fell to well below 3,000 microns. The total decrease can be estimated as about 700 microns at Lerwick and Rome and 1,000 microns at Oxford (the last being about one-third of the amount prevailing before the approach of the cold core).

Case B

The second case developed in January 1960 over North America. On 2 January (Fig. 7) the middle stratospheric circulation had its normal aspect, with a warm high over Alaska and a cold trough east of Greenland. The second and third weeks of the month were marked, however, by (i) the warming and intensification of the Alaskan high (see

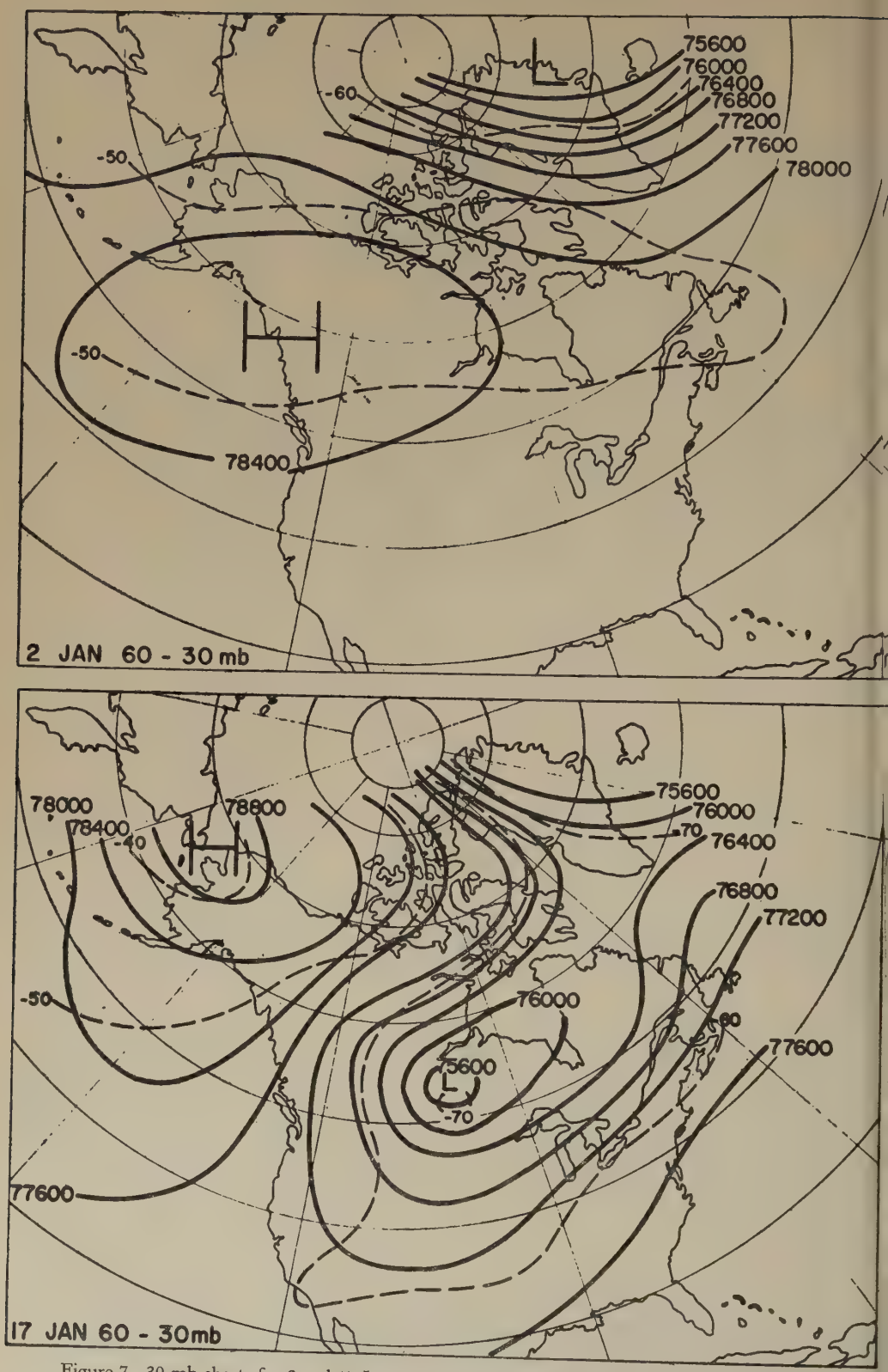


Figure 7. 30-mb charts for 2 and 17 January 1960, over North America, uniform with Fig. 2.

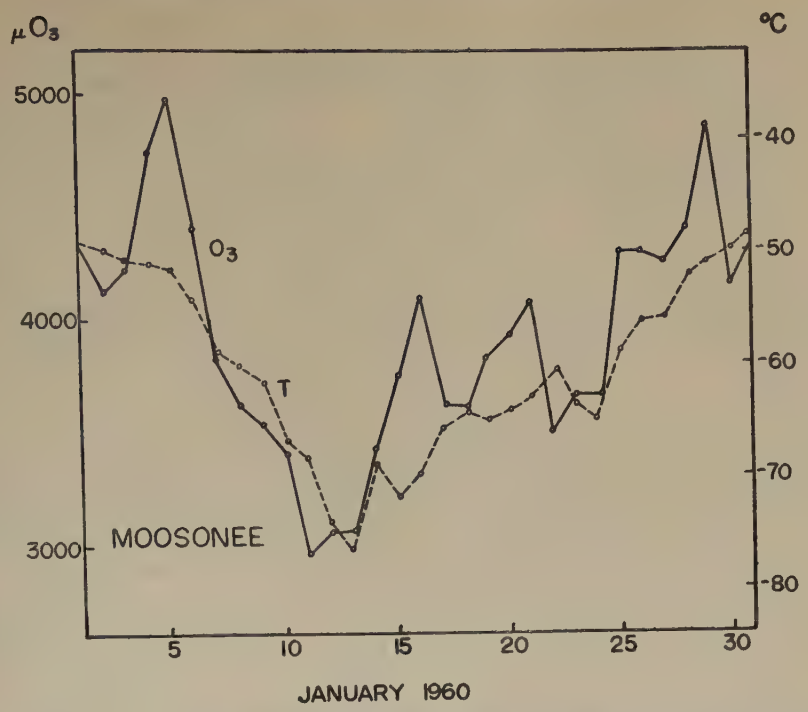


Figure 8. Time sections of 30-mb temperature (deg C, dashed line) and total ozone (microns, solid line) at Moosonee, Ontario, January 1960. The curves show the profound ozone and temperature changes, almost in phase, associated with the passage of the cold low shown on Fig. 7 (for 17 January).

Fig. 7, 17 January) and (ii) the deepening of a cold trough across Hudson Bay, with the creation by 10 January of a closed cold vortex which moved southwest to Saskatchewan by 17 January and into the western plains of the United States by 19 January. Thereafter it filled up slowly. The centre of the stratospheric low passed northwest of Moosonee, Ontario, on 12-13 January.

The consequences are shown in Fig. 8, which presents a time-section of 30-mb temperatures and total ozone at Moosonee for January 1960. The approach of the cold low was accompanied by a progressive fall of total ozone, from about 4,980 microns on 5 January to about 3,000 microns on 11-13 January, associated, as in *Case A*, with the temperature minimum. With the recovery of temperature after the low had passed away southwestward ozone also recovered to a more characteristic value.

4. The correspondence between total ozone and temperature variations

The case studies discussed, strongly support the hypothesis that middle stratospheric perturbations affect total ozone amounts in the manner predicted. An attempt will now be made to see whether the observed correspondence between ozone and middle stratospheric temperature variations is quantitatively of the right order of magnitude. Daily vertical ozone profiles were, of course, unavailable during the period of investigation, and it was necessary to devise a method based on standard *umkehr* profiles.

Fig. 1 shows standard vertical profiles of ozone mixing ratio, together with the vertical potential temperature profiles for a winter isothermal stratosphere. It is immediately clear that the profiles of the two elements can be matched almost perfectly by the use of a suitable scaling factor. By analogy, it is hence possible to find an analytical form for the vertical ozone profile. We write

$$\psi = a\theta + b \quad . \quad . \quad . \quad . \quad . \quad . \quad (1$$

where ψ = ozone mixing ratio, and θ = potential temperature. Hence

$$\psi = aTP^* + b \quad . \quad . \quad . \quad . \quad . \quad (2$$

where T = temperature, and $P^* = (1{,}000\ P^{-1})^{0\cdot 277}$, and

$$\sigma = \int_{P_1}^{P_2} (aTP^* + b)\, dP \quad . \quad . \quad . \quad . \quad (3$$

where σ is the ozone between the two pressure levels P_1 and P_2. The integration of Eq. (3 is especially simple for an isothermal atmosphere, in which case

$$\sigma = [1{\cdot}404\ aTPP^* + bP]_{P_1}^{P_2} \quad . \quad . \quad . \quad . \quad (4$$

The coefficients in Eq. (4) were determined from the standard ozone curve for mediur concentration for two layers, one set for the layer between 75 mb and 10 mb and anothe set for the layer between 250 mb and 75 mb. These were then applied to temperature (adjusted to noon local time) at Moosonee for January 1960; the 30-mb temperatur was used for the upper layer and the 200-mb temperature for the lower layer, but th pressure at the base of the lower layer was allowed to vary with the tropopause. Th computed ozone values were then correlated with the observed values. The *upper laye* (*30-mb temperature*) and total ozone gave a linear correlation coefficient of $r = 0{\cdot}85$ The *lower layer* (*200-mb temperature*) gave a linear correlation coefficient of $r = 0{\cdot}4$ and for a partial correlation coefficient the non-significant value of $R = 0{\cdot}23$. In spit of the crudity of the assumption concerning the ozone profile, this method therefor explains 72 per cent of the observed variance of total ozone in terms of middle stratospheri temperatures.

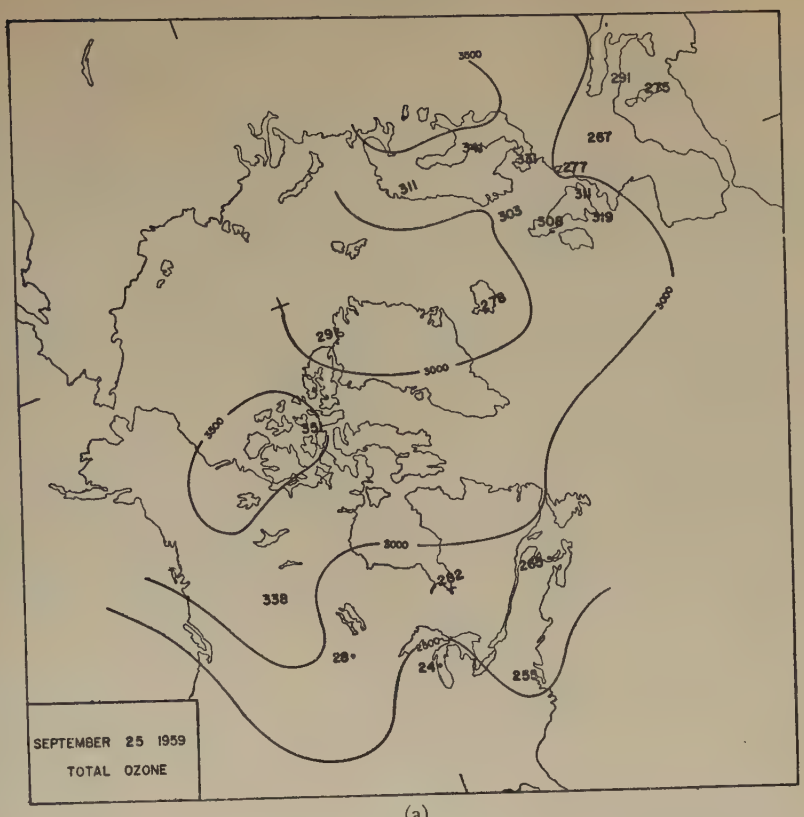
The lower stratospheric effect appears to be negligible at Moosonee, primarily becaus it is overshadowed by the larger activity in the middle stratosphere. It may also be note that it is difficult to filter out disturbances whose periods are only two or three times th ozone reporting interval of twenty-four hours. The phasing of observation times i particularly important at that lower limit of the time scale.

The middle stratospheric disturbances, generally having periods of over 10 days, will not suffer from the observational interval, and the phasing of data is not a problem. For these longer periods one would, however, expect non-conservative factors (radiative and chemical) to assume greater importance. An initial correspondence between the anomalies of total ozone and 30-mb temperature in the polar night systems can be accounted for by the effect of vertical motion in the presence of large vertical gradients of ozone mixing ratio and potential temperature at middle stratospheric levels. In addition, if (i) the basic horizontal distributions of the two parameters are similar, or the differences in the distributions are of a lower order of magnitude than the anomalies; and (ii) non-conservative effects do not strongly alter the anomaly structure, then subsequent horizontal advection and vertical motions, regardless of how they combine, will tend to maintain correspondence between the anomalies.

5. Hemispheric ozone patterns

It has been demonstrated that large changes in the middle stratospheric temperature patterns are accompanied by large ozone changes. Some further speculation on hemispheric ozone distributions is now presented.

During the early autumn, the middle stratosphere is essentially barotropic, with no horizontal temperature contrasts (Hare 1960b); thereafter, temperatures fall progressively in the polar region, reaching — 70C at 30 mb in November and — 80C in December; at the same times, the average temperatures in middle latitudes at 30 mb are about — 55C.



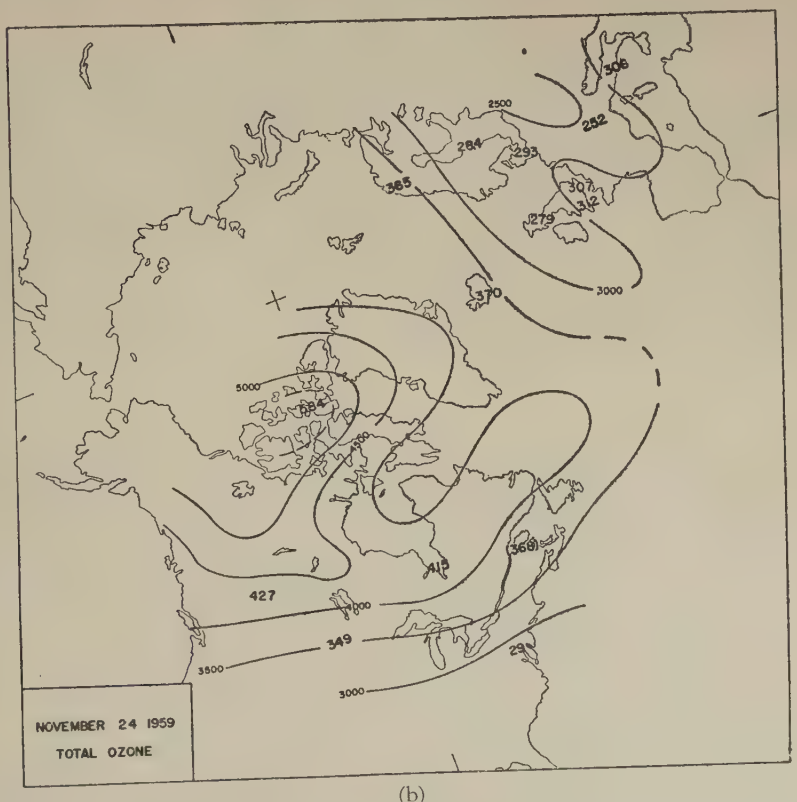


Figure 9 (a) and (b). Synoptic charts of total ozone concentrations (microns) for 25 September and 24 November 1959. Observed values are plotted in terms of microns (hundreds of microns where data are doubtful). The isopleths were drawn by the technique described in the text.

The mean latitudinal distribution of total ozone (Götz 1951) appears to follow a simila pattern, with amounts near 3,000 μ, and little gradient from mid-latitudes northwar in the autumn. Thereafter, amounts fall in the polar regions to near 2,000 μ in Decembe whereas in middle latitudes they remain constant or rise slowly. The hemispheric latitu dinal distributions of ozone and stratospheric temperature are thus similar, and it remain to discuss the dominant anomalies superimposed on the zonal pattern.

The major asymmetric feature of the polar-night stratospheric circulation is th Aleutian warm anticyclone. This warm zone has been related by Boville (1960) to th great intensity and persistence of tropospheric cyclogenesis in the Aleutian area. Durin October 1958, in association with hemispheric weather adjustments, there was intens cyclogenesis in that area. The vertical circulation associated with these events extende high into the less stable and slightly baroclinic middle stratosphere and created an anoma lous warm centre with temperatures at 25 mb near the — 40C mark over the Bering Sea, i contrast to temperatures near — 65C at similar latitudes in the Atlantic sector*. Unfor tunately ozone observations were not available to the writers for the Alaskan area, and th effects could not be examined in the warm zone source region. However, in Novembe 1958 the major part of the warm zone surged north-eastward into the Arctic Islands at time when lunar ozone observations were available from Resolute, North West Territories The three observations on 24, 25 and 26 November showed values *double* the expecte seasonal value and confirm the profound influence of the warm zones on the ozone distribu tion. These effects are shown schematically by the ozone distributions on Fig. 9 (a) and (b). The early autumn pattern is shown on the chart for 25 September (Fig. 9 (a)) and the warm zone effect by the one for 24 November (Fig. 9 (b)). The basis for these charts is described in the next paragraph. Similar phenomena occur in the Atlantic cyclogenetic sector; however, activity there is not as conservative in position nor as consistently intense as the Pacific counterpart, and Atlantic warm stratospheric anticyclones appear to be infrequent.

Highly tentative ozone distribution charts were made for the 1958-1959 year by entering five consecutive daily values for each station on a single chart : the centre day coincided with the dates of the 25-mb chart series. The similarity of the ozone distributions to the synoptic charts, together with the previously mentioned experiments, permitted the construction of ozone patterns. The isotherms of the 200-mb chart (1200Z) and the 25-mb chart were each converted to total ozone increments, using the potential temperature relationship of Eq. (4). These derived ozone patterns were then superimposed on the plotted chart and relaxed subjectively to fit the observed data. The distribution lines on Figs. 9 and 10 were drawn in this manner. Similar charts were drawn throughout the year, and there seems little doubt that all major perturbations of the polar-night vortex were accompanied by ozone anomalies of the expected sign.

Conclusions

The authors have throughout employed a synoptic-dynamic approach, and have avoided the physical problems of radiation and chemical activity. Although these are unquestionably fundamental factors determining the general distributions of both temperatures and ozone in the middle stratosphere, it appears permissible, as a first approximation, to neglect them when one considers synoptic-scale changes. The distributions in the tropics and near the tropical tropopause have been omitted at this stage as they require a different synoptic approach.

The above results, when combined with those previously presented by Allington, Boville and Hare (1960) justify the following conclusions :

(i) Total ozone in high latitudes in winter shows large variations ascribable to the passage of centres of high and low ozone moving with the phase velocities of the waves

* Arcas rocket winds at Churchill, Manitoba, recently communicated by Dr. W. L. Godson of the Canadian Meteorological Branch, suggest that a similar north-west thermal gradient reached up to about 50 km in October 1959.

the polar-night westerlies, low ozone accompanying the cold troughs and high ozone e warm ridges at levels above 100 mb (16 km).

(ii) This relation holds good even for closed lows that break away from the polar-ght westerlies. In such break-away systems, the ozone variations due to their passage e large enough to be clearly visible in the midst of oscillations due to the Reed-Normand fect in the Ferrel westerlies.

(iii) The results favour the dynamic activity of the polar-night vortex as a major gent in creating the large-scale anomalies in the total ozone distributions of middle and igh latitudes, a hypothesis similar to those previously proposed by Wexler and Moreland 1958) and Godson (1960).

Acknowledgments

The Research reported in this paper was supported by Air Force Cambridge Research Laboratories, Geophysics Research Directorate, U.S.A.F., Contract 19 (604)-3865 and, n the later stages of the work, by the Defence Research Board of Canada. Special thanks re due to Mr. Trevor Harwood of Defence Research Board for his interest and support.

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551.55 : 551.5

The flux of water vapour due to the mean winds and the convergence of this flux over the Northern Hemisphere in January and July

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(Manuscript received 11 October 1960)

SUMMARY

Charts are presented showing the flux of water vapour due to the mean winds, based on the mean distribution of water vapour and the mean geostrophic winds, in January and July, over an area comprising most of the Northern Hemisphere.

There is evidence that this flux approximates fairly well to the total flux of water vapour, but the divergence of this flux does not always approximate so well to the divergence of the total flux. Nevertheless, comparison with independent estimates of evaporation minus precipitation over the oceans suggests that the two divergences are at least of the same order of magnitude, and further comparison with published charts of the divergence of the total annual flux of water vapour suggests some degree of detailed correspondence.

1. The flux of water vapour

At any level in the atmosphere the product of the horizontal mass flow and the humidity per unit mass gives the horizontal flux of water vapour at that level. Taking simultaneous observations of humidity and wind at a given place from the surface upwards to a height above which the humidity content is negligibly small, vertical integration of the horizontal flux gives the total flux of water vapour above that place. From a time sequence of similar integrations the total flux of water vapour can be averaged over a period of time.

If instead of forming the time average of the product of humidity content and mass flow to obtain the average total flux, we take the time average of the humidity content multiplied by the time average of the mass flow, this flux will be referred to, for brevity, as the ' M-flux '*; the consequent neglect of eddy flux is discussed in Section 4.

Over most of the world the monthly average water-vapour content per unit area above the earth's surface and above the levels 850, 700 and 500 mb respectively, has been charted by Bannon and Steele (1960). Except in low latitudes and near the earth's surface the average observed winds averaged over a month are closely approximated by the average geostrophic winds, and the latter are readily derived from the monthly mean contour heights of the various isobaric surfaces. Taking the product of mean geostrophic wind and mean water-vapour content, layer by layer, (with special provision for the layer near the earth's surface where friction has a significant influence), the geostrophic M-flux of water vapour can be computed. The present investigation, which was initiated by Bannon, is concerned with this M-flux, and with the convergence of the M-flux.

2. Computation of the flux of water vapour due to the mean winds

The M-flux of water vapour may be represented by the vector $\mathbf{M}$ where

$$\mathbf{M} = (1/g) \int_0^{\bar{p}_0} \bar{q}\, \bar{\mathbf{V}}\, dp; \qquad (1)$$

the superscript bar denotes an average with respect to time (in this case over the month of January – or July – for the 5 years 1951-1955); g denotes the acceleration due to gravity; $\bar{p}_0$ the mean surface pressure; $\bar{q}$ the mean specific humidity and $\bar{\mathbf{V}}$ the mean wind; dp denotes an element of pressure-difference measured along a vertical axis.

* What is here termed the ' M-flux ' was called by Priestley (1949) the ' advective flux.'

To find $\overline{\mathbf{V}}$, mean charts were prepared for the period 1951-1955, January and July, to show

(a) mean-sea-level pressure

(b) height of the 850 mb surface

(c) height of the 700 mb surface

(d) height of the 500 mb surface

(e) height of the 300 mb surface

Sources of data are given in Appendix I.

From charts (b), (c), (d) and (e) were derived the mean winds at the respective levels, and from chart (a) the mean wind at about 950 mb. The surface wind was deduced as follows.

Over the sea, Sheppard (1949) has shown that, north of latitude 25°N, the angle between the surface wind and the surface geostrophic wind, derived from the sea-level isobars, is 10° or less, so here the mean surface wind over the sea was taken as having the same direction as the mean sea-level isobars of chart (a). Sheppard (1949) has also given a table from which can be deduced a proportionality between the surface wind speed over the sea and the surface geostrophic wind speed, for various latitudes between 5° and 55°, so here the mean surface geostrophic wind speeds were reduced in accordance with Sheppard's figures to obtain the mean surface wind speeds. North of latitude 55°, the ratio was assumed to be the same as at latitude 55°.

Sea areas covered by ice to the extent of making navigation impossible were reckoned as land, ice areas being defined as in the *Ice Atlas of the Northern Hemisphere* (1946).

Over land, the following rough rule was adopted after consulting Marshall (1954) and Shaw (1936): the mean surface wind direction was taken as backed by 30° from the mean sea-level isobars, and the mean surface wind speeds were taken as 30 per cent in January (35 per cent in July) of the respective mean surface geostrophic wind speeds.

The mean water-vapour content was derived from the charts by Bannon and Steele (1960) for the appropriate period, for the layers surface-850 mb, 850-700 mb, 700-500 mb and above 500 mb. These values were used in computing $\mathbf{M}$ in the following way. The average specific humidity $\bar{q}$ falls off rapidly with decreasing pressure, and it was found that the ratios of the values of $\bar{q}$ at 950 mb and 850 mb to the surface value (near 1,000 mb) are approximately the same at all seasons, except well inland over the continents in winter. The surface, 950 mb and 850 mb winds were therefore weighted according to the average ratios of the specific humidity at these levels, then the vector sum of these weighted winds was multiplied by the average water-vapour content of the layer, to give the transfer vector for the layer. Details are given in Appendix II. Similarly, the winds at the base and top of the layers 850-700 mb, 700-500 mb and 500-300 mb were weighted according to the average distribution of water vapour in the layer. (See Appendix II). The vector total of the contributions from all these layers made up the flux-vector $\mathbf{M}$. In detail, the (a) west to east and (b) south to north components of $\mathbf{M}$ were separately computed, (a) across 5° segments of meridian at 10° intervals of longitude, between 75° and 20°N, and (b) across 10° segments of parallels of latitude at 5° intervals of latitude from 20°N to 70°N, and across 20° segments of the parallel at 75°N. The flow across a segment was derived directly from the difference in contour-height (or mean sea-level pressure) between the two ends of the segment. The integrations in the vertical were evaluated by scalar additions for each component. The south to north component of flux across a segment of latitude was thus directly computed and the corresponding west to east component at the mid-point of the segment of latitude was taken as the average of the components across the four adjacent segments of meridian. It was then a simple step to calculate the direction and magnitude of the resultant flux-vector $\mathbf{M}$.

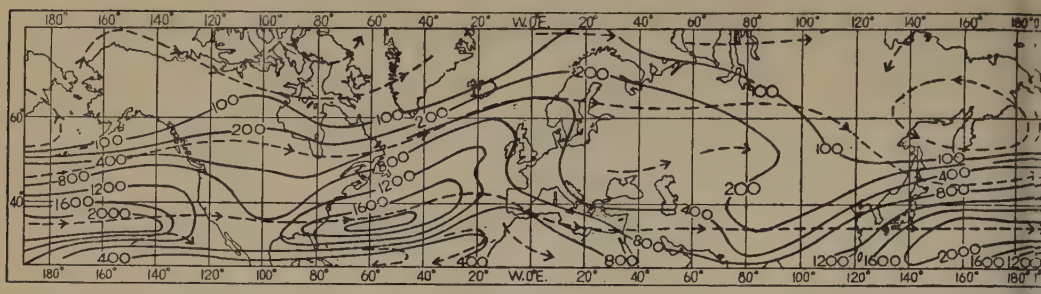

Figure 1. The flux of water vapour due to the mean winds in January (1951-55) : magnitudes shown by full lines with values in g cm^{-1} sec^{-1}, directions shown by pecked lines with arrows.

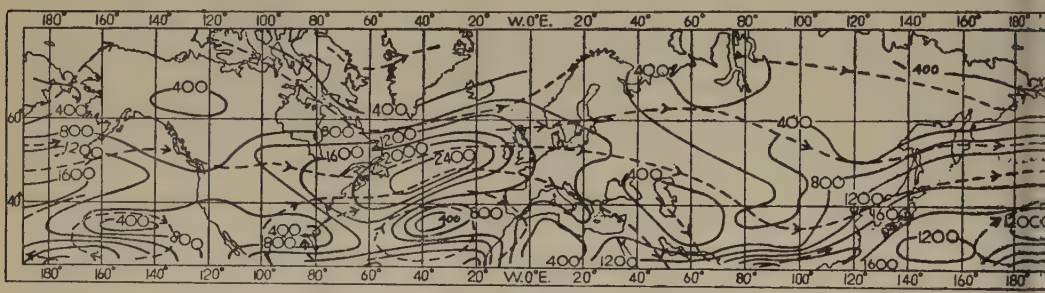

Figure 2. The flux of water vapour due to the mean winds in July (1951-55) : magnitudes shown by full lines with values in g cm^{-1} sec^{-1}, directions shown by pecked lines with arrows.

3. The flux of water vapour due to the mean winds in January and July, over the northern hemisphere

The results of the computations have been charted. Isopleths of the magnitude of the M-flux vector, and directions of the flux have been drawn, as shown, for January (Fig. 1) and for July (Fig. 2). The interval between isopleth-values is 400 g cm sec^{-1} except in January (Fig. 1) where two additional isopleths help to fill the large, otherwise blank, area in the north.

4. Significance of the flux of water vapour due to the mean winds

As explained in Section 1, the M-flux which is charted in Figs. 1 and 2 represents an approximation to the total flux by neglecting the eddy flux. In Eq. (1), considering the integrand at the level p, and resolving the wind vector into components, u directed west to east and v directed south to north, the M-flux vector at this level is

$$(\overline{q}\,\overline{u},\ \overline{q}\,\overline{v}) = (\overline{qu},\ \overline{qv}) - (r_{q,u}\,\sigma_q\,\sigma_u,\ r_{q,v}\,\sigma_q\,\sigma_v) \quad . \quad . \quad . \quad (2)$$

where $(\overline{qu},\ \overline{qv})$ denotes the total flux at this level; $r_{q,u}$, $r_{q,v}$ denote the correlation coefficients between q and u, and between q and v, respectively and σ_q, σ_u, σ_v denote the relevant standard deviations. If the wind distribution is approximately circular, $\sigma_u = \sigma_v = \sigma/\sqrt{2}$ where σ denotes the standard vector deviation, and Eq. (2) can be re-written

$$(\overline{q}\,\overline{u},\ \overline{q}\,\overline{v}) = (\overline{qu},\ \overline{qv}) - \{\sigma_q\,\sigma/\sqrt{2}\}\,(r_{q,u},\ r_{q,v}) \quad . \quad . \quad . \quad (3)$$

Thus in the case of a circular wind distribution, at any level the direction of the eddy-flux vector is determined by the ratio of the correlation coefficients $r_{q,u} : r_{q,v}$, while the magnitude of the eddy-flux is proportional to σ_q, σ and $\{(r_{q,u})^2 + (r_{q,v})^2\}^{\frac{1}{2}}$.

The eddy-flux gives rise, in the general case, whether the wind distribution is circular or not, to component integrals, west to east and south to north, respectively :

$$g^{-1}\int_0^{\bar{p}_0}\sigma_q\,\sigma_u\,r_{q,u}\,dp, \quad g^{-1}\int_0^{\bar{p}_0}\sigma_q\,\sigma_v\,r_{q,v}\,dp \qquad . \quad . \quad . \quad (4)$$

Values of the standard vector deviation of wind have been charted over the world by Brooks *et al.* (1950), and show considerable variability in the horizontal, in the vertical, and with time of year. Notably high values of σ occur in winter over and to west of the British Isles. Data for σ_q and $r_{q,u}$, $r_{q,v}$ are lacking on a wide scale, but orders of magnitude over the British Isles are indicated by values for Fazakerley in Table 1, based on two observations daily in January 1951 and 1952, at convenient isobaric levels. In addition, Table 1 gives the component standard deviations of wind σ_u, σ_v, and the products denoted by

$$g^{-1}\,\overline{(q'\,u')} \equiv g^{-1}\,\sigma_q\,\sigma_u\,r_{q,u}$$

$$g^{-1}\,\overline{(q'\,v')} \equiv g^{-1}\,\sigma_q\,\sigma_v\,r_{q,v}.$$

The peculiar but convenient units following Hutchings (1957) viz., g (cm sec 100 mb)$^{-1}$, are well suited to carrying out the integrations in Eq. (4) in units of g cm sec^{-1}. The vertical integrals of the eddy-flux are also given, the west to east component on the left side of Table 1, the south to north components on the right; for comparison, similar data are given for the M-flux as defined in Eq. (1).

It is evident that the M-flux vector in Table 1 is of much greater magnitude than the eddy-flux vector, but it is also noteworthy that the meridional components of the two vectors happen to be opposite in sign and of nearly equal magnitude.

TABLE 1. COMPONENTS OF EDDY-FLUX AND M-FLUX OF WATER VAPOUR, CORRELATIONS OF WIND COMPONENTS WITH HUMIDITY, AND STANDARD DEVIATIONS, AT FAZAKERLEY, JANUARY 1951, 1952

	(West to east)							(South to north)	
	M-flux	Eddy flux						Eddy flux	M-flux
	$g\left\{\begin{matrix}\text{cm}\\100\,\text{mb}\\\text{sec}\end{matrix}\right\}^{-1}$	$g\left\{\begin{matrix}\text{cm}\\100\,\text{mb}\\\text{sec}\end{matrix}\right\}^{-1}$	m sec^{-1}	g (10^3 g)$^{-1}$	m sec^{-1}			$g\left\{\begin{matrix}\text{cm}\\100\,\text{mb}\\\text{sec}\end{matrix}\right\}^{-1}$	$g\left\{\begin{matrix}\text{cm}\\100\,\text{mb}\\\text{sec}\end{matrix}\right\}^{-1}$
Level (mb)	$g^{-1}\,(\bar{q}\,\bar{u})$	$g^{-1}\,\overline{(q'\,u')}$	$r_{q,u}$	σ_u	σ_q	σ_v	$r_{q,v}$	$g^{-1}\,\overline{(q'\,v')}$	$g^{-1}\,(\bar{q}\,\bar{v})$
1,000	135	22	0·45	4·9	1·00	3·7	0·18	7	71
950	286	38	0·47	8·2	0·95	7·4	0·11	8	48
850	279	50	0·51	9·3	1·03	7·7	0·03	2	− 18
700	155	18	0·22	10·3	0·79	8·8	0·14	10	− 30
500	65	12	0·28	13·7	0·31	14·2	0·06	3	− 16
Vertical integral	938	149	←	units : g cm^{-1} sec^{-1}			→	30	− 36

Values of total flux of water vapour, M-flux and (by subtraction) eddy flux, were computed directly for the stations given in Tables 2 and 3 for certain winter and summer months.

At most of these stations, the M-flux is seen to be a fairly good approximation to the total flux, in each component, with the eddy flux often an order of magnitude smaller, but particularly in winter the meridional component of eddy flux in some cases exceeds the corresponding component of the M-flux, notably at Brest and Ship *J*. Benton and Estoque (1954) found that over U.S.A. the eddy flux contributes 42 per cent of the total meridional flux of water vapour, but only 6 per cent of the total zonal flux.

TABLE 2. EASTWARD AND NORTHWARD COMPONENTS OF FLUX OF WATER VAPOUR DURING A WINTER MONTH JANUARY 1951 (g cm^{-1} sec^{-1}).

Month	Station	Eastward component of			Northward component of		
		total flux	M-flux	eddy flux	total flux	M-flux	eddy flux
January 1951	Lerwick	346	277	69	545	501	44
	Stornoway	802	628	174	529	479	50
	Leuchars	923	817	106	163	190	− 27
	Aldergrove	1,103	1,034	69	284	266	18
	Fazakerley	1,130	1,003	127	128	139	− 11
	Downham Mkt.	1,119	914	204	142	128	14
	Valentia	1,423	1,331	92	131	48	83
	Larkhill	1,221	1,028	193	213	73	140
	Camborne	1,372	1,243	129	187	74	113
	Ship *J*	2,091	2,085	6	433	101	332
	Sola	204	116	88	407	349	58
	Jever	1,049	960	89	281	198	83
	De Bilt	1,280	1,092	188	378	335	43
	Trappes	1,314	1,213	101	202	275	− 73
	Brest	1,467	1,310	157	132	− 1	133
	Bordeaux	938	829	109	− 107	− 115	8
	Malta	825	877	− 52	− 127	− 157	30
	Gibraltar	1,089	1,019	70	− 445	− 501	56
	Kagoshima	858	884	− 26	− 87	− 156	69
January 1953	Charleston	1,412	1,496	− 84	954	468	486

Upper-air observations in January 1951 were available four times a day at some of these stations.

TABLE 3. EASTWARD AND NORTHWARD COMPONENTS OF FLUX OF WATER VAPOUR DURING A SUMMER MONTH (g cm^{-1} sec^{-1})

Month	Station	Eastward component of			Northward component of		
		total flux	M-flux	eddy flux	total flux	M-flux	eddy flux
June 1955	Lerwick	470	342	128	407	156	251
	Stornoway	428	315	113	634	489	145
	Leuchars	669	463	206	356	299	57
	Aldergrove	649	497	152	729	673	56
	Fazakerley	817	687	130	603	525	78
	Hemsby	909	702	207	207	108	99
	Valentia	1,067	997	70	1,185	972	213
	Crawley	818	763	55	381	348	33
	Camborne	861	897	− 36	825	703	122
	Malta	1,136	1,073	63	− 310	− 380	70
	Gibraltar	290	394	− 104	465	385	80
July 1951	Kagoshima	2,392	2,170	222	1,160	1,088	72
July 1952	Charleston	383	237	146	694	678	16
July 1953	Charleston	429	173	256	445	339	106

Hutchings (1957) found that the eddy flux made only a small contribution to the total flux over southern England in the summer of 1954.

Flohn and Oeckel (1956) investigated water-vapour flux during the summer rains over Japan and Korea (July 1952), and found the magnitude of the eddy-flux vector to be about one-sixth of the magnitude of the total flux vector.

It is considered therefore that the M-flux of water-vapour charted in Figs. 1 and 2 obably represents a fairly good approximation to the total flux.

The total flux of water vapour is of direct interest in two respects; firstly, in its relation precipitation and evaporation, and secondly in its role in the energy balance as a conveyor latent heat.

Main features of the flux of water vapour due to the mean winds over the northern hemisphere

The fall-off in the values of the water-vapour content with height usually outweighs he increase of wind with height, and in consequence, the direction of the M-flux vectors ears a close relation to the direction of the geostrophic winds at a low level*, around 50 mb.

Over most of the hemisphere both in winter and in summer, the content of water apour above the surface shows a considerable gradient towards higher values in the outh, except over North Africa and Tibet; in winter the content of water vapour is igher over central oceanic areas than over central continental areas in the same latitude. n winter, the M-flux of water vapour reaches a maximum over the Atlantic and Pacific Oceans, round latitudes 32 to 35°N, in the region of strong geostrophic winds at 850 mb; arther south, towards the light winds of the anticyclonic belt, the M-flux decreases to ery small values. In summer the magnitude of the M-flux shows greater complexity, vith minima around latitude 35°N over the Atlantic and over the eastern Pacific, corres- onding with areas of light geostrophic winds in the anticyclonic belt and, in the case of the astern Pacific, relatively low water-vapour content. There are, however, maxima over he western Pacific where the geostrophic winds are significant, and also in middle latitudes n the central Atlantic, where the water-vapour content is at least as high as over the adjacent ontinents and where the 850 mb winds are relatively strong.

Over the North American continent and adjacent oceans, Figs. 1 and 2 may be compared with the charts presented by Benton and Estoque (1954), bearing in mind that the latter harts refer to the total flux of water vapour in a single year (1949), for the months of anuary and August and for the seasons of winter (January, February, December) and summer. The authors point out that several of the months during 1949, notably January, August and November, exhibited unusual mean circulation patterns; nevertheless, their harts of total flux show a fair similarity to the M-flux in Figs. 1 and 2 over the relevant areas, particularly their chart for summer 1949 and Fig. 2. In January 1949 the maximum total flux was centred some 25 degrees west of the maximum M-flux in Fig. 1; over Tennessee, magnitudes of total flux were twice the M-flux values given in Fig. 1. Over the Pacific the maximum of total flux was centred 15 degrees north of the maximum of M-flux shown in Fig. 1.

Comparison may also be made with the findings of Flohn and Oeckel (1956), who investigated the flux of water vapour during the summer rains over Japan and Korea. At four stations (Wakkanai, Kimpo, Haneda and Kadena) they found the magnitude of eddy-flux to have been only about one-sixth of the magnitude of total flux. Fig. 2 confirms that south of latitude 32°N, the flux is towards the north or north-east but over northern Japan and South Korea, where Flohn and Oeckel found the total flux of water vapour directed towards the east or east-north-east and emanating from the continent of Asia, in July and August 1952, the M-flux shown in Fig. 2 suggests a more northeasterly direc- tion. Thus the M-flux over northern Japan is directed from the East China Sea and the South China Sea. This suggests that the moisture supplying the summer rainfall over Japan may be in part derived from the southerly monsoonal winds of the China Seas.

Over eastern China, however, Fig. 2 shows the M-flux directed from the land towards the ocean. This is in harmony with the findings of Hsieh Yi-ping and Tai Wu-ja (1959) for a selected synoptic case, viz., 11-20 July 1957.

* Hsieh Yi-ping and Tai Wu-ja (1959) make a similar comment on their findings for eastern China in 11-20 July 1957.

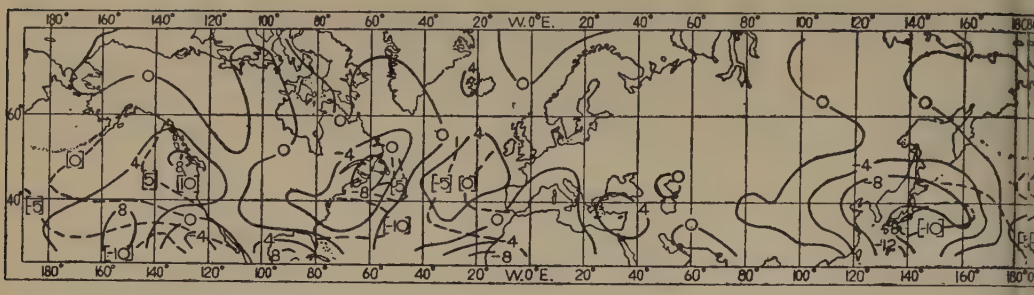

Figure 3. Convergence (positive) and divergence (negative) of the flux of water vapour due to the mea winds in January (1951-55), shown by full lines, with values in g cm^{-2}. Pecked lines with values in squa brackets denote $\frac{1}{3}(P - E)$ derived from Jacobs's seasonal values of Evaporation minus Precipitation for th northern oceans.

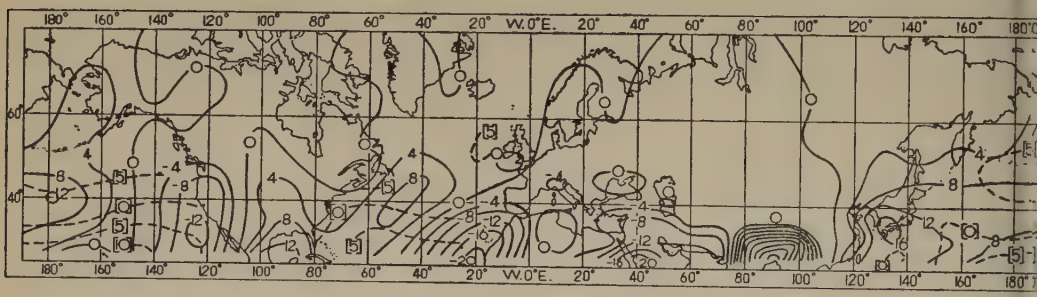

Figure 4. Convergence (positive) and divergence (negative) of the flux of water-vapour due to the mea winds in July (1951-55), shown by full lines with values in g cm^{-2}. Pecked lines with values in square bracket denote $\frac{1}{3}(P - E)$ derived from Jacobs's seasonal values of Evaporation minus Precipitation for the northern oceans.

6. Convergence of the flux of water vapour due to the mean winds

Computations were also made of the net M-flux of water vapour into each small area bounded by 5°-lengths of meridian, and 10°-lengths of parallel for latitudes 25-70°N (20°-lengths of parallel at 75°N). These areas were nearly rectangular and the corresponding convergence into each was evaluated in units of g cm^2 month^{-1}.

Although the results showed a consistent broad pattern, there was considerable scatter. This scatter arose from a lack of smoothness in the values of wind flow read off the contour charts, and to a lesser extent in the values of water-vapour content. The effects of high ground may also have led to inconsistencies in some adjacent values of the convergence. Accordingly, an average was taken of the values of convergence over each set of four adjacent 'rectangles' having a common corner at the point to which the average was assigned. This gave a smooth field of the convergence (positive) and divergence (negative) of the M-flux, as shown in Fig. 3, January, and Fig. 4, July.

7. Significance of the convergence of total flux of water vapour

If over some period of time the *total* horizontal flux of water vapour is known into, and out of, the sides of a vertical cylinder of unit horizontal cross-section with its base at the earth's surface, continuity requires that the net inflow, or convergence, is balanced

y the sum of

(i) the increase in water-vapour content from the beginning to the end of the period;

(ii) the increase in condensed form;

(iii) the outflow of water vapour and condensed water vapour through the top f the cylinder, and

(iv) the net outflow of water vapour through the base of the cylinder at the earth's urface, outwards in the condensed form of precipitation, dew, etc., and inwards in the orm of evaporation from the surface.

Orders of magnitude may be estimated as follows. Hutchings (1957) found that ver southern England in the summer of 1954, item (i) contributed about 1 g cm^{-2} and hat (ii) was quite negligible. Consider item (iii), and denote by w the upward vertical elocity at the ' lid ' of the cylinder, say, 300 mb. Then the outflow through the lid may e taken as $\overline{pqw} \simeq \bar{p}\,(\bar{q}\,\bar{w} + \overline{q'w'})$ per unit area, where a superscript bar denotes a mean alue with respect to time, and a primed quantity denotes a deviation from the mean. An order of magnitude for the mean vertical velocity in various latitudes may be taken rom the analysis by Tucker (1959) of meridional circulations based on the sector 160°W o 40°E. In summer, the maximum mean vertical velocities at high levels are rather less han 1 cm sec^{-1} downwards around latitude 35 to 40°N and 0·5 cm sec^{-1} upwards around atitude 45°N. The mean specific humidity at 300 mb ranges from about 0·1 to 0·8 g per 10^3 g. Taking then $\bar{w} \sim 1$ cm sec^{-1}, $\bar{q} \sim 0{\cdot}3$ g per 10^3 g, density $\bar{p} \sim 4{\cdot}6\,(10^{-4})$ g cm^{-3}, gives a mean vertical flux over the month of July around latitudes 35 to 40°N of order ·4 g cm^{-2}. A similar or smaller magnitude applies to January around latitude 25°N. f the vertical eddy contribution is of similar magnitude or smaller, item (iii) can be neglected.

The dominant item balancing the convergence of total flux is thus (iv), that is, precipitation and evaporation.

The preceding remarks applied to the convergence of the total flux, and it is necessary now to consider whether this may be replaced by the convergence of the M-flux as shown n Figs. 3 and 4.

8. Significance of the convergence of the flux of water vapour due to the mean winds

Although the M-flux of water vapour is generally the dominant component of the total flux, and may be regarded as a fairly good approximation to the total flux, it does not follow that the convergence of the M-flux is an equally good approximation to the convergence of the total flux. This is most easily seen in the extreme case of a locally non-divergent field of M-flux of large magnitude : the convergence of the total flux would be entirely given by the convergence of the small eddy flux. Benton and Estoque (1954) concluded from their computations of the water balance over the North American continent that when divergence fields are considered, the importance of eddy-transfer terms is far greater than their magnitudes might suggest.

On the other hand, Hutchings (1957) showed that over southern England in the summer of 1954, the convergence of eddy flux of water vapour made only a small contribution to the convergence of total flux.

It was considered desirable to investigate the convergence of the eddy flux of water vapour over the whole British Isles area in a winter month (January 1951) and in a summer month (June 1955). Accordingly, values of the total flux, M-flux and eddy flux of water vapour at the British Isles network of upper-air stations were taken to define linear vector fields over triangles with vertices at adjacent stations. A formula essentially equivalent to that quoted by Sheppard (1949) was used to determine objectively the divergence over each triangle, allowance being made for the earth's curvature. Results are given in Table 4. Another method of computing the divergence was also employed, taking a grid of ' squares '

and calculating the net flux out of each square, based on subjectively-drawn isopleth of the field of flux. The square-grid method gave similar results to the triangulation method.

TABLE 4. CONVERGENCE OF FLUX OF WATER VAPOUR OVER THE BRITISH ISLES IN A WINTER AND A SUMMER MONTH

Month	Triangle, with vertices at	Convergence ($g\ cm^{-1}$) of Total flux	M-flux	Eddy flux
January 1951	Liverpool, Larkhill, Downham Mkt.	22	9	13
	Liverpool, Camborne, Larkhill	19	13	5
	Liverpool, Leuchars, Downham Mkt.	7	14	− 7
	Aldergrove*, Stornoway, Lerwick	12	0	12
	Aldergrove*, Leuchars, Liverpool	5	10	− 5
	Aldergrove*, Camborne, Liverpool	4	4	0
	Aldergrove*, Camborne, Valentia†	8	3	5
	Aldergrove*, Lerwick, Leuchars	− 46	− 30	− 16
June 1955	Liverpool, Crawley, Hemsby	6	− 11	17
	Liverpool, Camborne, Crawley	0	3	− 3
	Liverpool, Leuchars, Hemsby	16	22	− 6
	Aldergrove, Stornoway, Lerwick	− 4	4	− 8
	Aldergrove, Leuchars, Liverpool	10	9	1
	Aldergrove, Camborne, Liverpool	− 3	0	− 3
	Aldergrove, Camborne, Valentia†	37	29	8
	Aldergrove, Lerwick, Leuchars	− 17	− 1	− 16

It will be noted from Table 4 that the convergence of the eddy flux was usually small, often of the order of the standard error (an estimate, on the lines of Hutchings (1957) treatment, suggests a standard error here† of roughly 2·5 $g\ cm^{-2}$). By this standard, the convergence of the eddy flux is not always negligible, for example, in June 1955, over the triangle Liverpool, Crawley, Hemsby, where the convergence of the eddy flux considerably exceeded the corresponding convergence of the M-flux. Comparison of the values of convergence of total flux in Table 4 with measurements of rainfall at places within the appropriate triangles suggests no very close correspondence such as might have been expected at least in January when evaporation is small. Even allowing for the unrepresentativeness of individual measurements of rainfall these differences suggest that topographical features, in particular, coastlines and hills, tend to put the assumption of a linear field of flux in considerable error, particularly in the northern triangle, Aldergrove*, Lerwick, Leuchars, in January 1951, where the convergence of total flux was substantially negative although rainfall must have considerably exceeded evaporation.

Nevertheless, the variability of the convergence of eddy flux in Table 4 appears to support the views of Benton and Estoque (1954) applying to the North American continent as relevant to the British Isles area also : it is unsafe to regard the convergence of the M-flux of water vapour as a uniformly good approximation to the convergence of the total flux.

* Many observations were missing at Aldergrove in January 1951 and interpolated values may have badly affected the calculated convergence in triangles marked with an asterisk. However, the triangle Aldergrove, Lerwick, Leuchars gives a peculiar value for the convergence in June 1955 also, perhaps suggesting a strong topographical effect.

† In the case of the triangle involving Valentia, where only one observation, (at 14 hr), was available daily, the error probably exceeds 4·5 $g\ cm^{-2}$.

A comparison can be made over the oceans in middle and low latitudes from the ublished estimates of $E - P$, i.e., evaporation minus precipitation, as calculated by Jacobs 950) in a quite independent way. Both estimates and correspondences can only be ough, and so it was considered adequate to divide by 3 Jacobs's seasonal 3-monthly values, ntred on January and July. Isopleths based on Jacobs's charts are accordingly shown pecked lines on Figs. 3 and 4, with values in square brackets (all signs reversed to give $- E$). These may be compared with the isopleths of convergence here, which are own on the same Figures as full lines, with values attached.

The magnitudes are similar.

This is perhaps all that should be expected in the circumstances. Privett (1960) akes a similar point in comparing his investigations of energy transfer, in particular, vaporation over the Southern Oceans, with notable investigations by Albrecht (1949); acobs (1942), (1951); Budyko (1955) : these four sets of results agree in establishing the orrect order of magnitude, but they do not agree in detail. Precipitation amounts over ne oceans have also been rather speculative hitherto : values over the North Atlantic Ocean may have to be reconsidered in the light of an ingenious statistical investigation by Tucker (1961).

Figs. 3 and 4 may be compared with a recent determination of the horizontal diver- ence of the vertically integrated *total* annual flux of water vapour for the year 1950 over he Northern Hemisphere, as presented by Peixoto (1959).

Bearing in mind the possibility of some variation in the annual pattern from year to ear, Peixoto's annual values for 1950 of the total divergence of water-vapour flux show ufficient broad agreement with Figs. 3 and 4 to suggest that the M-flux provides an mportant contribution to the total flux, even for considerations of divergence, but there re also some large discrepancies which suggest that the contribution of the eddy-flux o the divergence is sometimes important.

Acknowledgment

This paper is published by permission of the Director-General of the Meteorological Office.

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Appendix I

Sources of data

The charts of average mean-sea-level pressure, and heights of the 850, 700, 500 and 300 mb levels fo the months of January and July over the period 1951-55 were constructed using the following data :

(a) Mean-sea-level pressure

The maps in *Die Grosswetterlagen Mitteleuropas* gave partial coverage of the area.

' Climat ' data, i.e., monthly means received telegraphically from various selected stations, were availabl in the Meteorological Office.

Values for various points in the North Pacific were taken from daily charts given in the Daily Seri Synoptic Weather Maps.

(b) 850 mb

' Climat ' temperature data were available in the Meteorological Office.

Heights were calculated for a grid of points over the polar regions, U.S.S.R., Atlantic and Pacific Ocean using the charts of average mean-sea-level pressure in (a), average surface temperatures and average tempera tures at 850 mb evaluated mainly from *Climat* temperature data. For Russian stations, 850 mb temperature extracted by Mr. G. A. Tunnel for the period 1948-52 were used. Average surface temperatures were derive from the following publications : *Arctic Atlas*, 1952, Washington Hydrographic Office, Publication 573 *Klima der U.S.S.R.*, 1932, Teil 1, Leningrad; *Monthly Meteorological Charts of the Atlantic and Pacif Oceans*, Meteorological Office, London, M.O. 483, 484, 578.

The height of the 850 mb surface is not very sensitive to the mean temperature of the layer underneath and this method of computing the height of the 850 mb surface makes full use of surface pressures ove areas for which upper-air data are few.

(c) The charts for the remaining levels were constructed from the individual charts prepared in th Meteorological Office from ' Climat ' temperature data, taking account at 500 mb of the charts published in *Die Grosswetterlagen Mitteleuropas*, which cover part of the area.

Appendix II

Components of flow of moisture across segments of latitude and longitude

Except well inland over the continents in winter and near the extreme west coast of America, the ratio of values of water-vapour content at 950 and 850 mb to the surface values are approximately the same in summer and winter. Accordingly the following weighting factors were used in computing the M-flux o water vapour.

(*a*) *January – Surface to 850 mb.*

Over land in areas 75° to 40°N over North America including the Atlantic coast but excepting the extreme west coast, over Eurasia north of 45°N and east of 20°E and also north of 40°N and east of 40°E, the above ratios were about unity, so that water-vapour contents of the layer, surface to 850 mb, was multiplied by $\frac{1}{3}(u_s + u_g + u_{85})$, $\frac{1}{3}(v_s + v_g + v_{85})$ for the west to east and south to north components respectively, where suffix *g* denotes the geostrophic values derived from the mean-sea-level isobars, suffix *s* denotes the derived surface wind components, and suffix 85 denotes the geostrophic values derived from the 850 mb contour lines.

For all other land areas, sea areas and ice covered sea, the ratios of water-vapour content at surface, 950 and 850 mb were about 1·6 : 1·4 : 1, so the water vapour content of the layer was multiplied by

$$(0{\cdot}4\,u_s + 0{\cdot}35\,u_g + 0{\cdot}25\,u_{85}), \quad (0{\cdot}4\,v_s + 0{\cdot}35\,v_g + 0{\cdot}25\,v_{85}).$$

850 to 700 mb, all areas

The water-vapour content of the layer was multiplied by $(\frac{2}{3}u_{85} + \frac{1}{3}u_{70})$, $(\frac{2}{3}v_{85} + \frac{1}{3}v_{70})$ where u_{pp}, v_{pp} denote the west to east and south to north components of the geostrophic wind at the isobaric level 10 (*pp*) mb.

700 to 500 mb, all areas

The water-vapour content of the layer was multiplied by $(\frac{3}{4}u_{70} + \frac{1}{4}u_{50})$, $(\frac{3}{4}v_{70} + \frac{1}{4}v_{50})$.

500 to 300 mb, all areas

The water vapour content of the layer was multiplied by $(0{\cdot}86\,u_{50} + 0{\cdot}14\,u_{30})$, $(0{\cdot}86\,v_{50} + 0{\cdot}14\,v_{30})$.

(*b*) *July*

The same weighting factors of the various wind components were everywhere the same as for January except in the lowest layers.

Surface to 850 mb, all areas

The water vapour content of the layer was multiplied by $(0{\cdot}4\,u_s + 0{\cdot}35\,u_g + 0{\cdot}25\,u_{85})$, $(0{\cdot}4\,v_s + 0{\cdot}35\,v_g + 0{\cdot}25\,v_{85})$.

551.515.72 : 551.515.8 : 551.576.2

Thermal structure in the lower layers of anticyclones

By J. FINDLATER
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(Manuscript received 13 February 1961)

SUMMARY

The amount of stratocumulus cloud in easterly anticyclonic airstreams over the British Isles is shown to be related to the pressure and temperature at the base of anticyclonic inversions. These two parameters are not conservative. Fluctuations occur, often in a periodic manner, as shallow but distinct cold and warm areas circulate in the lower layers of anticyclones. These cold and warm areas originate from successive pulses of cold and warm air on the flanks of anticyclones but retain their identity after the fronts separating them have decayed and been omitted from routine analyses. Approximate dimensions and forms of significant temperature patterns are deduced.

1. INTRODUCTION

During periods of anticyclonic weather the amount of stratocumulus cloud reported by a station may vary between nil and total cover, even when the geostrophic trajectories over a 24 hr period, the direction and speed of approach of the air, the month and time of day are all similar. The differing cloud regimes in such cases are undoubtedly due to varying lapse rates of temperature and moisture in the layer of air between the surface and the anticyclonic inversion. Study of the lapse rates of temperature in this zone led to two parameters being selected as having some importance in determining the liability of stratocumulus cloud formation. These were the pressure and temperature at the base of an anticyclonic inversion. For the purposes of this investigation isothermal layers have been treated as inversions.

2. THE AMOUNT OF STRATOCUMULUS CLOUD

The data used have been the surface observations from Leuchars, Ringway and St. Eval for the month of April in each of the five years 1953-1957 inclusive, on occasions when the geostrophic trajectory of the air during the previous 24 hr displayed anticyclonic curvature, and no fronts lay within 200 naut. mi of a station at the time of observation.

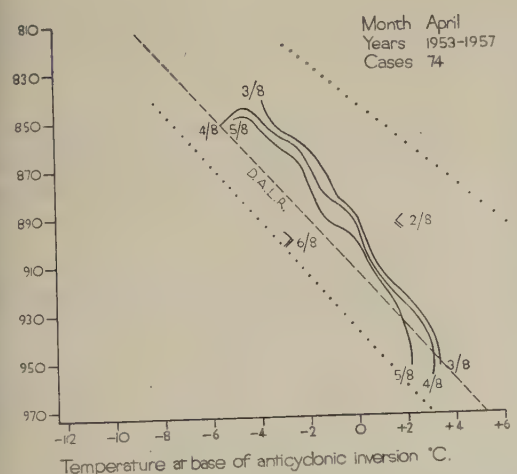

Figure 1. 1200 GMT

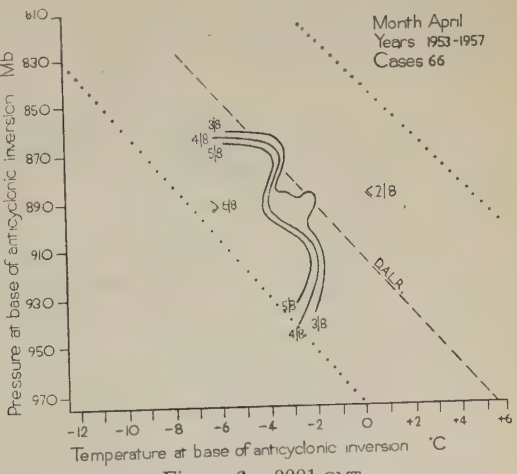

Figure 2. 0001 GMT

Figures 1 and 2. Mean isopleths of amount of stratocumulus for Leuchars, Ringway and St. Eval with geostrophic winds E'ly or less than 5 kt. The dotted lines contain the scatter of observations.

Upper-air data were obtained from the soundings nearest to the stations concerned, i.e., from Leuchars, Liverpool and Camborne.

To relate cloud reports to the pressure and temperature at the base of the anticyclonic inversion, diagrams of the type shown in Figs. 1 and 2 have been prepared by plotting each report of stratocumulus amount, including nil reports, in a position appropriate to the characteristics of the inversion base. Block means, calculated over squares of 20 mb × 2°C, have been used to draw isonephs. A large number of such diagrams have been prepared for different combinations of stations, air masses and times, yet nearly all of them show a clear division between clear ($\leqslant 2/8$) and cloudy ($\geqslant 6/8$) conditions. The most marked boundary is shown in Fig. 1 where reports from the three stations at 1200 GMT have been combined for occasions when the geostrophic wind had an easterly component or was less than 5 kt.

The isopleths on Fig. 1 lie near the dry adiabatic from the mean temperature of the North Sea in April, suggesting that stratocumulus formed when convection from the sea surface extended up to the inversion base. The comparable pattern for 0001 GMT is shown in Fig. 2, and although the boundary remains well defined, the reason for its change of form is not obvious. The sharpness of these boundaries is incompatible with the variability of the amount of stratocumulus cloud in anticyclones unless the pressure and temperature at the base of the anticyclonic inversion are also liable to frequent change. Accordingly, the next stage in the investigation was to examine the temporal changes in the height and temperature of anticyclonic inversions.

3. Analyses of time cross-sections

Nine prolonged periods of anticyclonic conditions over the British Isles were chosen, not necessarily in April, but in the winter or spring months of the years 1954-60 inclusive. All these periods were at least nine days in length and airflow from between NE and SE predominated. Time cross-sections showing isotherms in the layer from the surface to 750 mb were prepared for one or more upper-air stations in the British Isles, and in some cases for the Ocean Weather Stations at 59°N 19°W and 52° 30′N 20°W. Soundings at 12 hr intervals were used as published in the *Daily Aerological Record*; the special reports at levels of significant change of lapse rate allowed isotherms to be drawn accurately near inversion levels. A study of the cross-sections revealed the following points of interest :

(*i*) Variation of the height and temperature of the inversion base, and even discontinuities, were too erratic to allow the use of extrapolation methods in making 12 hr forecasts of these parameters. Periods in excess of one day when the pressure and temperature at the inversion base remained sensibly constant were rare.

(*ii*) Shallow, but well-defined, warm and cold centres were frequently apparent near the level of the inversion. These areas moved at a speed close to that of the wind at their level, and often appeared in series, recurring at intervals of between 36 hr and 48 hr.

(*iii*) The cold areas were often centred around peaks in the inversion level, but warm areas always lay above the main inversion and were usually located above dips in the inversion level. On some occasions a secondary inversion lay above a warm area.

(*iv*) The cold and warm centres occurred at any hour of the day and night and appeared over land and sea.

(*v*) Large amounts of stratocumulus cloud were generally associated with cold areas, although diurnal effects overland were noted.

(*vi*) The shallow warm and cold areas displayed depth/length ratios of between 1/600 and 1/800 in most cases.

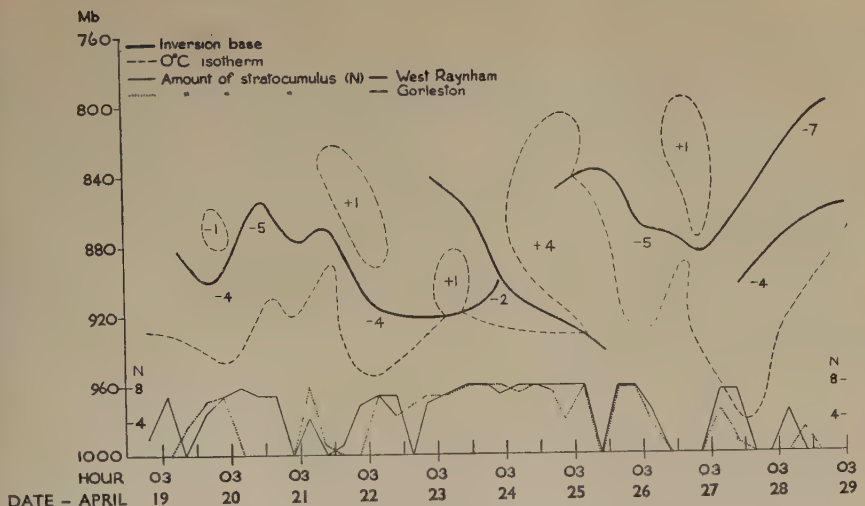

Figure 3. Time cross-section through Hemsby from 0300 GMT, 19 April 1954, to 0300 GMT, 29 April 1954. Maximum and minimum temperatures of warm and cold areas, respectively, are shown in °C. Short vertical lines indicate the times of upper-air soundings.

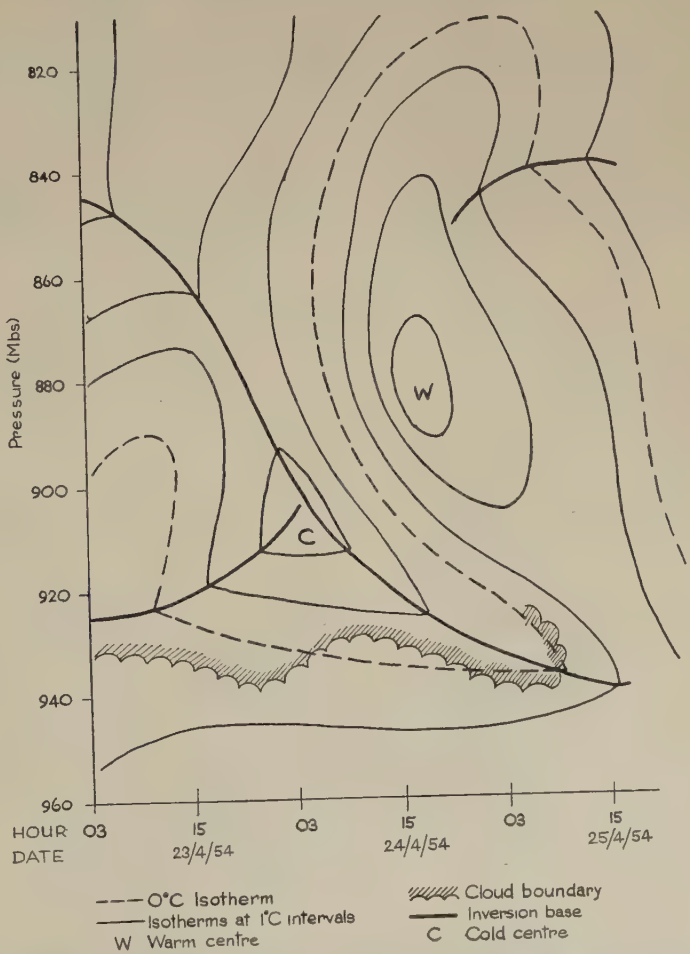

Figure 4. Time cross-section through Hemsby from 0300 GMT 23 April 1954 to 1500 GMT 25 April 1954 showing detail of warm and cold areas near the inversion level. Short vertical lines indicate the times of upper-air soundings.

An example of one of the time cross-sections, simplified by showing only the isotherm for 0°C is shown in Fig. 3. In this diagram many of the dips in the level of the 0°C isotherm are reflected as peaks in the amount of stratocumulus, but not all such cross-sections showed this strong association. It may also be noted in this case that the low-level cold centres recurred at intervals of about 48 hours, and several other such diagrams which have been prepared for different periods and stations show periodicities of this type to be fairly common in anticyclonic airstreams.

On time cross-sections, 121 cold centres have been analysed and of these 109 have 4/8 or more stratocumulus cloud under the centres, whereas only 12 have less than 4/8. Thus the cold centres were nearly always cloudy whereas warm areas were associated with either clear or cloudy skies, depending on the relative heights of the convective condensation level and the inversion base.

Warm centres, though usually occurring between cold centres, at times nearly overlay them. An example of one detailed pattern is shown in Fig. 4. Wet-bulb potential temperature analyses have revealed an increase with height of a few degrees Celsius through the inversion base in most cases when these centres were present. The horizontal temperature changes are usually about 4°C/12 hr, as typified by Fig. 4, but changes of 10°C have been noted, also an extreme value of 16°C. The existence of these patterns as travelling features has been confirmed by space cross-sections at 12 hr intervals, on which significant areas have been tracked across the British Isles. An example of a space cross-section along an E'ly anticyclonic airstream, showing two cold areas and their associated cloud layers, is shown in Fig. 5. This cross-section has been drawn using data from seven upper-air soundings and from horizontal traverses by an aircraft of the Meteorological Research Flight. The cloudy areas moved from east to west parallel to the line of the cross-section and a little drizzle fell from the leading feature. Plan-view analyses of the cloud structure on this occasion showed that 8/8 stratocumulus cloud lay to the north whilst over France the air was cloudless. The boundary, in the horizontal, of the cloud layer was of wavelike form and the line of the selected cross-section lay across this pattern.

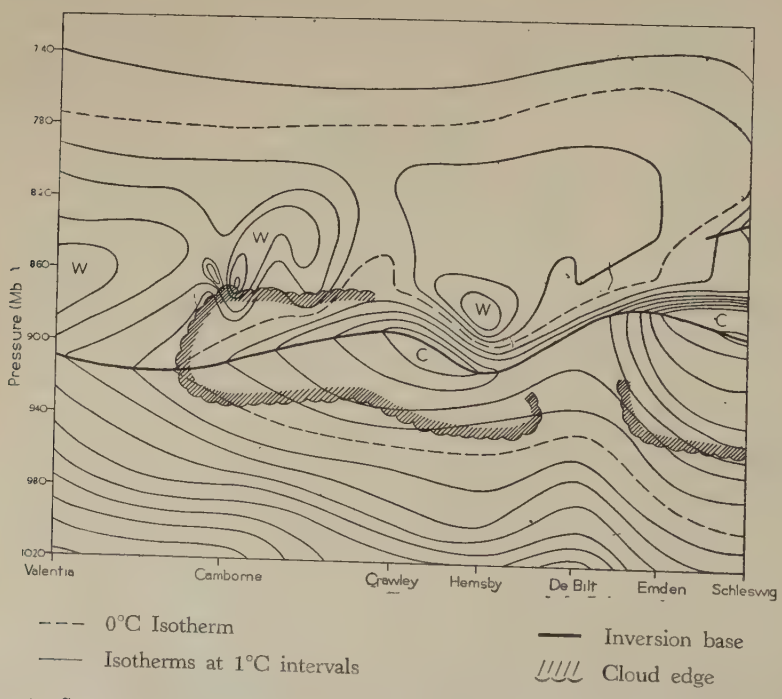

Figure 5. Space cross-section from Valentia to Schleswig at 1200 GMT on 17 March 1959.

Time cross-sections were constructed during several cyclonic situations for comparison ith those of anticyclones, but the temperature patterns showed little more than ridges nd troughs. Even when transient inversions were present no distinct cold and warm ntres appeared.

4. Thermal patterns

On time and space cross-sections it has been noted that on some occasions the arm centres lie between the cold centres, and on other occasions directly above them. ositions intermediate between these two extremes occur frequently. Most of the zones f maximum horizontal temperature change, as shown on cross-sections, conform to ne of the four types shown in Fig. 6. The 121 cold features with an associated warm rea which have been analysed can be allocated to the different types of pattern as follows :

Type 1	32	Type 3	28
Type 2	33	Type 4	15
	Unclassified	13	

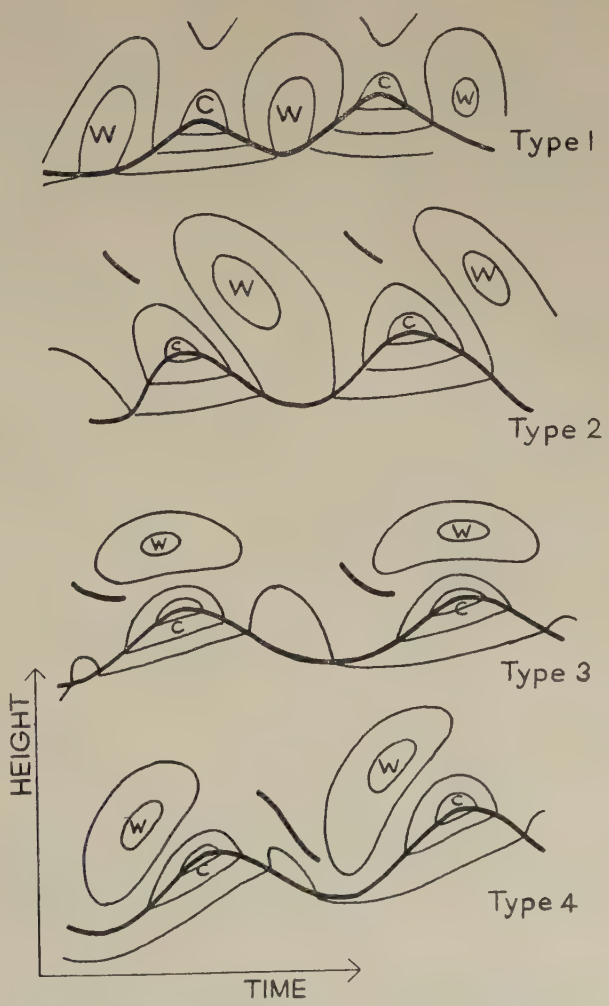

Figure 6. Four types of temperature patterns found on time cross-sections.

5. Origin and movement of thermal patterns

A distinctive feature of many of the cold areas has been a marked peak in the lev of the anticyclonic inversion, while under warm zones a dip in the level has usually bee evident. To find the sources of these air-masses geostrophic tracking for lengthy perio has been carried out when the air at about 600 m was in close association with a cold warm centre. 36 centres have been tracked in this way, and after periods of from fi to ten days, 16 cold areas were traced to high latitudes, and 15 warm areas to low latitude The remaining five areas, three cold and two warm, could not be tracked through wide area near the Alps and their origins were undetermined.

It would appear, therefore, that these temperature patterns are primarily due t advection, though they suffer modification on passage. Nearly all the cold areas whe tracked back geostrophically for several days were in cold air-masses whose fronts ha been omitted from routine analyses as they decayed in the fringes of anticyclones, ofte about five days before their arrival over the British Isles. Warm areas, however, althoug coming from low latitudes were rarely bounded by a warm front during the period f which they were tracked.

Of the cases studied, a particular one having a cold area at the inversion level whic became entrained into a warm sector, before being brought over the British Isles t Liverpool, was the most interesting though it may not be unique. The track of this a is shown in Fig. 7. Time cross-sections drawn for a selection of upper air stations alon the track have confirmed the passage of a cold area within 24 hr of the time expecte by geostrophic displacement. Trajectories of the air reaching Camborne have shown tha the air had a similar, though slightly more S'ly, track than that arriving at Liverpoo Upper-air geostrophic trajectories at 700 and 500 mb have also been drawn from Britis stations and the temperatures on these trajectories deduced by interpolation from nearb

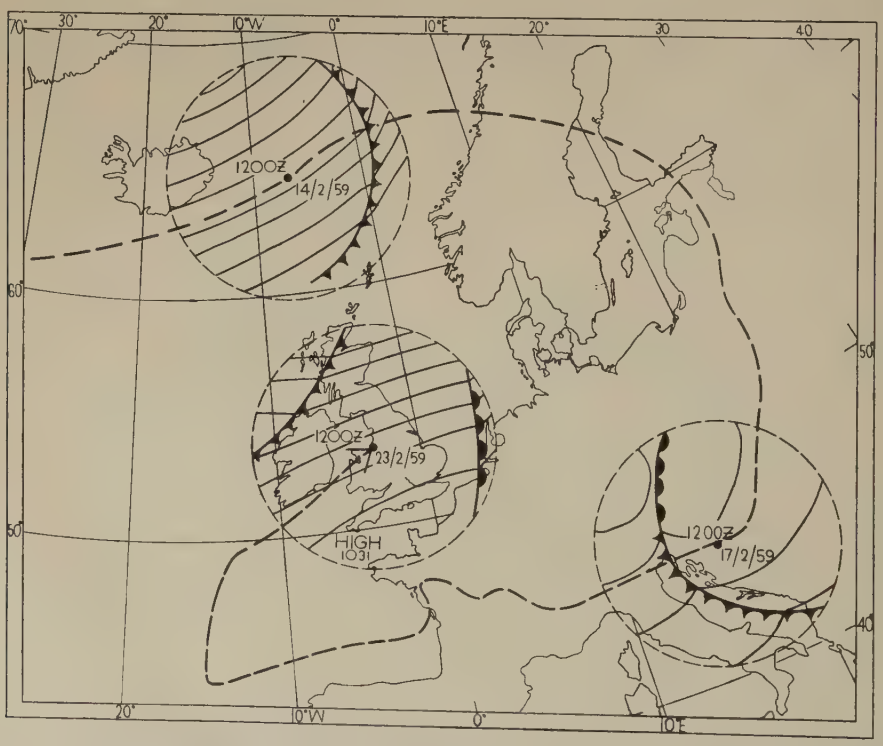

Figure 7. Geostrophic trajectory of the air, at about 600 m, which arrived over Liverpool at 1200 GMT on 23 February 1959.

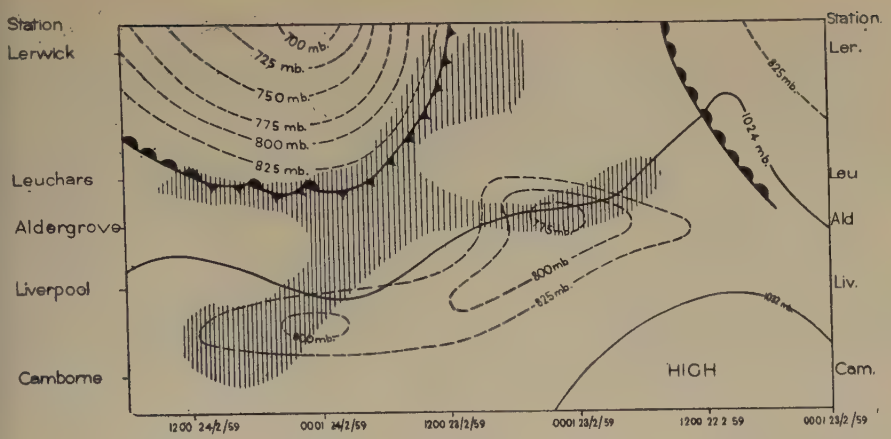

Figure 8. Time chart through a line from Lerwick to Camborne, showing the rain area (hatched) associated with a cold front and domes of cold air in the warm sector. Isobars represented by dashed lines indicate the top of the cold air.

upper-air stations. Although the trajectories at different levels had widely differing origins the temperature on each trajectory changed only slowly, at the rate of about 1°C per day and often less. During the period of one to two days trajectories at the same level were drawn from markedly different latitudes and it would appear that the temperature profiles recorded by upper air stations in the period studied were due largely to differential advection.

To clarify the vertical and horizontal extent of the mass of cold air whose track is shown in Fig. 7, space cross-sections along a line from Lerwick to Camborne (i.e., nearly normal to the airflow) have been drawn at 12 hr intervals and these have been used to construct the analysis shown in Fig. 8 – a representation of the changes in the topography of the cold air with time. The time period of $2\frac{1}{2}$ days corresponds to a run-of-wind of approximately 1400 naut. mi. The broken isobars showing the extent and depth of the cold air in the warm sector represent the pressure at the height of the inversion. In this inversion an increase with height of 4°C in the wet-bulb potential temperature took place. Surface observations at 6 hr intervals at stations within 50 naut. mi of the line of each cross-section have been used to insert the precipitation area, shown hatched, in its appropriate position on Fig. 8. Precipitation occurred close to the position of the cold dome in the warm sector at 0001 GMT on 23 Feb. 1959, and later merged with rain from the cold front. Later, as the front became stationary a separate area of precipitation appeared in association with a secondary cold dome at 0001 GMT on 24 Feb. 1959.

6. Horizontal extent of thermal patterns in anticyclones

The process of locating cold and warm areas by cross-section analysis is time-consuming and it was considered desirable to find whether analysis of charts of upper-air parameters could assist in finding and tracking significant sectors in anticyclones. The following parameters have been charted to examine the movement, horizontal extent, and relation to layer cloud, of shallow warm and cold areas near the inversion levels of anticyclones :

(*i*) Pressure and temperature at the base of anticyclonic inversions;

(*ii*) 850 mb temperature;

(*iii*) 1,000-850 mb thickness.

(a) Charts of pressure and temperature at the base of anticyclonic inversions

Since the cold areas are associated with rather higher inversions than the warm areas plan-view charts have been prepared for several periods to show the pressure and the potential temperature at the base of the anticyclonic inversion. The potential temperature was used in preference to actual temperature since the isotherm of the former appropriate to the sea temperature would be more readily indicative of areas likely to promote convection up to the level of the inversion.

On these charts well-marked warm and cold areas were evident and were frequently defined by potential temperature gradients of 8°C/100 naut. mi across their boundaries. The plan-view patterns of isotherms were most often in the form of deep troughs and ridges and occasionally as cut-off areas. A large variety of shapes and sizes was apparent but, in general, the length along the airstream and the breadth across were measurable

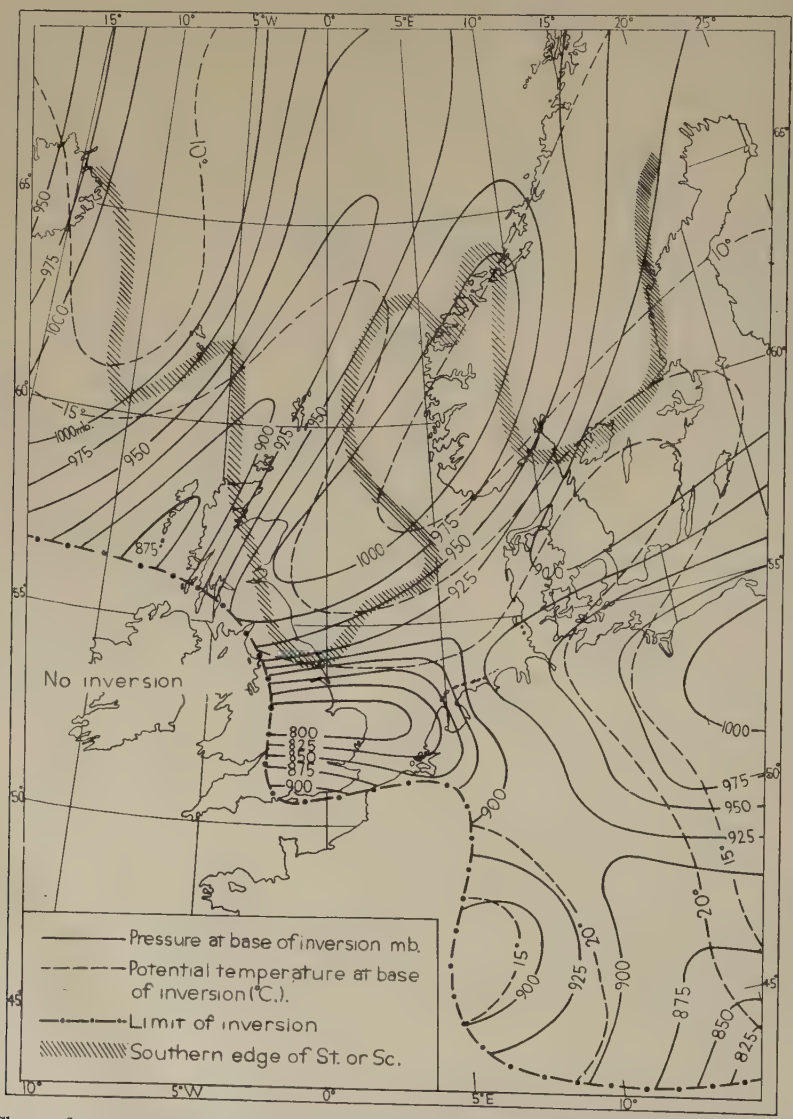
No inversion
Pressure at base of inversion mb.
Potential temperature at base of inversion (°C.).
Limit of inversion
Southern edge of St. or Sc.

Figure 9. Chart of pressure and potential temperature at the base of anticyclonic inversions at 2300 GMT on 7 October 1959. An anticyclone over Scandinavia maintained SE'ly winds over the North Sea and adjoining land areas.

hundreds of miles. With dimensions of about 450-600 m vertically and 200-300 naut. i (350-550 km) horizontally in the direction of the wind, the depth/length ratio of bout 1/700 in this synoptic situation was similar to that shown on time and space cross-ections. Fig. 9 shows a plan-view analysis for 2300 GMT on 7 October 1959. The spatial ariations of the pressure and temperature at the base of the inversion and the subsequent iability to variations in low cloud and visibility indicates the weakness of classifying air ccording to a 24 hr trajectory, at least insofar as aerodrome weather characteristics in nticyclonic conditions are concerned.

A time cross-section through De Bilt (06260) during the same period showed that he air leaving the Dutch coast, although cloudless, was inhomogeneous at levels above which heating from the sea surface could be expected to extend. These heterogeneities ere less marked than on their subsequent arrival at Leuchars. The movement of warm nd cold areas over the continent and the southern part of the North Sea between con-ecutive charts at 12 hr intervals was in good agreement with that expected from geostrophic lisplacement. Closed centres tended to change into troughs and ridges, and vice versa, etween consecutive charts. In the northern part of the North Sea, however, a marked eature of the pattern was a persistent trough of low potential temperature at the base of he inversion. This trough was associated with persistent low stratus on the east coast f Scotland, and occasional stratus over north-east England. Warm areas advected towards he cold trough tended to lose their identity on plan-view analyses, but were nevertheless uite evident as alternate warming and cooling on the time cross-section through Leuchars.

The pattern of stratus over the North Sea and adjacent land areas was determined y analysis of all available ship and land reports of which there were generally sufficient o define significant cloud zones. At Leuchars stratus cloud was a persistent feature and he cloud sheet appeared to be little affected by the movement of temperature systems bove. Periodic fluctuations in amount such as have been noted previously with strato-cumulus cloud were not apparent and on the few occasions when the low cloud dispersed here was no suggestion that warm features aloft had contributed to the clearance. Never-heless, there was a broadscale association between the persistent trough near east Scotland and the large cloud amounts in that area, as typified by the patterns shown in Fig. 9.

(*b*) *Charts of 850 mb temperature*

Charts of the temperature at 850 mb have been prepared for the same synoptic situation as for the charts described above. Tongues and pools of warm and cold air have appeared at the 850 mb level throughout the life and extent of the anticyclone. Temperature gradients of about 8°C/100 naut. mi were noted when fronts penetrated into or were engulfed by anticyclones, but even after the fronts had weakened and been omitted from routine analyses for several days temperature gradients of 4-6°C/100 naut. mi were occasionally evident. A serious disadvantage of this type of chart is that shallow cold features associated with an inversion at about 900-950 mb are not necessarily revealed by 850 mb temperature patterns. Several occasions have been noted where cold features at low level were disguised by warm features above them affecting the 850 mb level. This effect is particularly noticeable on some of the time cross-sections.

Two interesting features have emerged from a study of the 850 mb temperature charts. Firstly, the temperature gradients associated with weak cold fronts were occasion-ally well ahead (e.g., 250 naut. mi) of the surface positions of the cold fronts. Secondly, marked sinuosities in the temperature field at 850 mb are apparent on many occasions yet no such distortions appear in the surface positions of the fronts.

(*c*) *1,000-850 mb thickness charts*

Analysis of 1,000-850 mb thickness charts showed patterns rather similar to those of the temperature at 850 mb. Shallow warm and cold areas were again masked to some extent and their positions could not be determined with accuracy. In the case of the

series of temperature patterns illustrated in Fig. 3 the 1,000-850 mb thickness change were too small to be easily related to the shallow underlying features. Charts of thi 1,000-850 mb thickness, as with those of the temperature at 850 mb appeared to be c little value in identifying and tracking significant temperature fluctuations at the bas of the inversion.

7. Conclusions

The shallow and recurrent cold and warm areas which have been noticed to occu close to the level of anticyclonic inversions derive from successive pulses of cold and warm air of limited depth on the flanks of anticyclones. The fronts separating these shallow air masses are often omitted from routine analyses as they decay or change form wher approaching high-pressure areas. Nevertheless, the air masses may retain their identity when circulating in the anticyclones several days after such omissions. The cold and warm areas display depth/length ratios of approximately 1/700 and move at a speed close to that of the wind at their level.

Cold areas, after passing over an extensive water surface which can produce convection up to the inversion level are prone to the formation of stratocumulus cloud.

Acknowledgment

This paper is published by permission of the Director-General of the Meteorological Office.

551.501.81 : 551.578.71

Multi-wavelength radar reflectivity of hailstorms

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Imperial College, London

(Manuscript received 22 February 1961; in revised form 21 June 1961)

SUMMARY

Computations are made of the radar reflectivity Z_e of particles present in hailstorm updraughts in exponential size spectra. The reflectivity varies substantially according to the radar wavelength and according whether the particles are wet or dry. In severe storms a very strong echo may be observed aloft. Partly reference to a particular example it is shown that its magnitude at the shorter wavelengths $Z_e \approx 10^7\ \text{mm}^6\ \text{m}^{-3}$), and the variation of its intensity with wavelength, can reasonably be explained only the responsible particles are large dry hailstones of almost uniform size. The dryness and sorting of the ge stones is attributed to fall outside the updraught through a wind shear, and the decrease of echo intensity wer down to their re-entry into the updraught and wetting. It appears that the likelihood and size of hail the ground are indicated by echo intensity aloft at 3 cm, and less sensitively at 4·7 cm, but not at 10 cm. easurements of echo intensity at several wavelengths are helpful in the study of severe storms.

1. INTRODUCTION

Donaldson (1958, 1959, 1960) finds that the intensity of radar echoes from hailstorms remarkably high, the radar quantity Z_e reaching 10^7 and even $4 \times 10^7\ \text{mm}^6\ \text{m}^{-3}$. He has also found that the more severe a storm is, the greater is the height of the most intense echo and the amount by which its intensity exceeds that of the echo near the ground, where Z_e is limited to about $10^6\ \text{mm}^6\ \text{m}^{-3}$.

In this paper we try to explain these features, referring to calculations of the radar reflectivity of certain size spectra of hailstones (using new data) and to observations of particular storm made with radars of three wavelengths.

2. THE SIZE SPECTRA OF HAILSTONES

The size spectrum of the particles inside a hailstorm has not yet been measured, and so must be inferred from observations of the hail and rain which reaches the ground. Prohaska (1905) noticed that very large hail not infrequently falls on the right flank of a storm with hardly any rain, and this happened also in the particular storm to be discussed. In this example, the cumulonimbus anvil spread forward and to the left of the storm, and on this side the hail-size diminished until eventually at some distance there was only rain at the ground. Supposing that all the precipitation is grown in the strong updraught it appears that there is a tendency for the particles to become sorted according to size during their fall in the strong wind shear which typically accompanies severe hailstorms (Newton 1960; Dessens 1960). Consequently, the size spectrum of the particles which arrive on the ground differs from place to place; to infer a likely spectrum of particles within a storm cloud Ludlam and Macklin (1959) assembled values for the concentration of the largest stones which were observed at a variety of places in the path of the severe storm of 5 September 1958, near Horsham, England. They have since obtained similar values for the storm of 9 July 1959, near Wokingham, England; when all values are plotted, as in Fig. 1, they indicate a spectrum which has approximately the exponential form :

$$N_D = N_0\, e^{-\Lambda D} \qquad (1)$$

where N_D is the concentration of stones with diameters between D and $D + \Delta D$, $N_0 = 40\ \text{m}^{-3}\ \text{cm}^{-1}$ and $\Lambda = 2{\cdot}27\ \text{cm}^{-1}$.

* Work completed whilst on detached duty from the Geophysics Research Directorate, Air Force Cambridge Research Center, Bedford, Massachusetts (as National Science Foundation Senior Post-Doctoral Fellow).

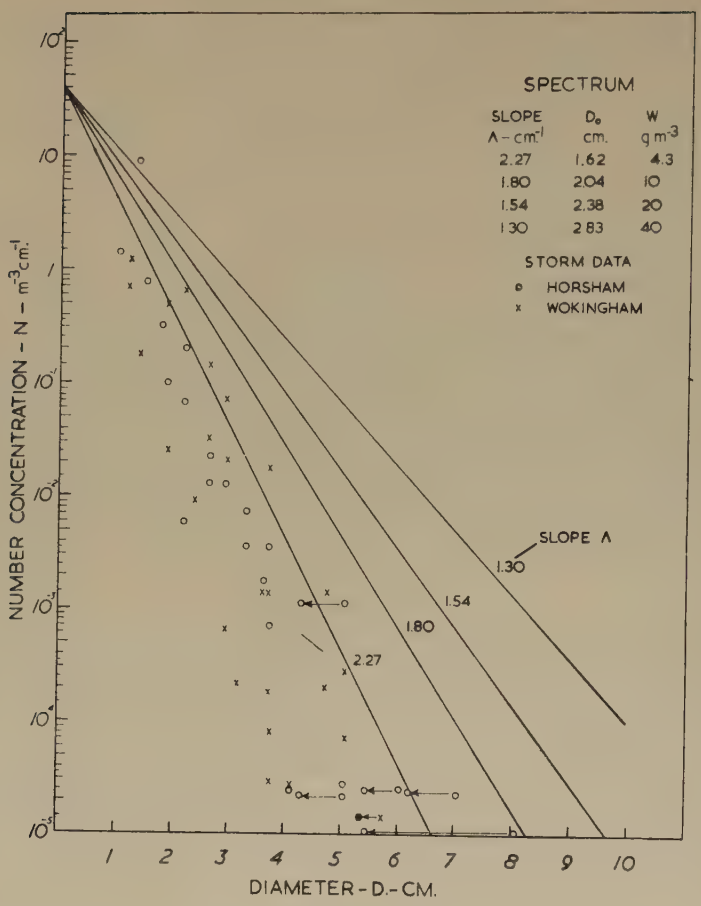

Figure 1. Concentrations in the air of the largest stones which fell at particular places in the Horsham and Wokingham storms (according to Ludlam and Macklin), and four exponential hailstone spectra used in the calculations referred to in the text. In the case of markedly aspherical stones, arrows lead to points representing the diameter of spherical stones of the same mass. The concentration plotted is that deduced for the — 10°C level, after assuming a melting in a downdraught of speed equal to the fallspeed of the stone.

Since similar laws have been found as approximations to the size spectra of particles in both rain (Marshall and Palmer 1948) and snow (Gunn and Marshall 1955) it is reasonable to use such a relation to compute the echo intensity which might be expected from a hailstorm.

With the values of N_0 and Λ quoted, the total (frozen) water content W is 4·3 g m^{-3} and R, the precipitation rate, is 278 mm hr^{-1}. As we are interested in the radar reflectivity of the most intense storms, we must consider that occasionally greater values of R have been observed in rainstorms, implying that higher values of precipitation concentration are possible, even up to the 35 g m^{-3} near the ground which corresponds to the world record rate of about 1,000 mm hr^{-1} for a fall of 5 min duration (Chomicz 1951; the rain is assumed to approach the ground at a mean speed of 8 m sec^{-1}). Accordingly, we have made calculations with the three other spectra shown in Fig. 1 and specified in Table 1. In these, N_0 is always 40 m^{-3} cm^{-1}, but the slope Λ decreases so that in the last spectrum W reaches 40 g m^{-3}, although we consider that such a concentration aloft is unlikely ever to be attained, since it would destroy the buoyancy corresponding to temperature excesses of more than 15°C (Ludlam 1958), which are not encountered even in adiabatic models of convective storms.

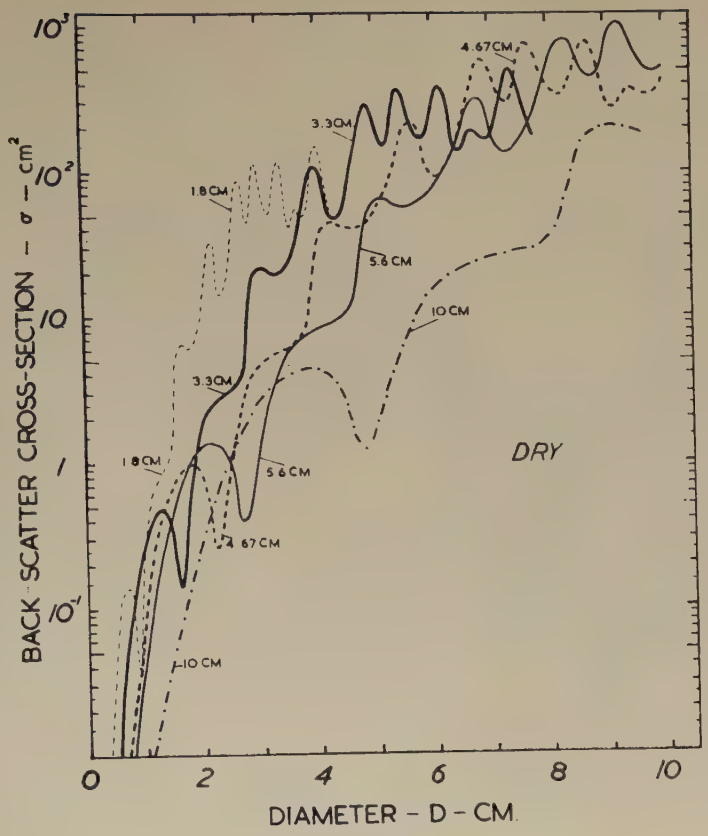

Figure 2 (a). Back-scatter cross-section σ of spheres of ice, with dry surfaces, at the indicated wavelengths.

TABLE 1. PARAMETERS OF SELECTED EXPONENTIAL HAIL SIZE SPECTRA

Spectrum	Λ cm^{-1}	Median volume diameter cm	W g m^{-3}	R mm hr^{-1}
1	2·27	1·62	4·3	278
2	1·80	2·04	10	731
3	1·54	2·38	20	1,580
4	1·30	2·83	40	3,440

(all have $N_0 = 40$ m^{-3} cm^{-1}).

3. THE RADAR CROSS-SECTION OF HAILSTONES

Values* of the back-scatter cross-section σ of spheres of ice and liquid water, for several commonly-used wavelengths, are given in Figs. 2 (a) and (b). For large spheres, $D > \lambda$, the ice values may exceed the water values by 10 db or even more.

* Values for ice are based on the work of Atlas *et al.* (1960), and those for water on the work of Aden (1951); strictly the latter's data apply only to a wavelength of 16·23 cm and a water temperature of 20°C, but they are accurate to within about 4 per cent for all wavelengths down to 3 cm, and temperatures down to 0°C, for $D/\lambda \geqslant 0{\cdot}3$. We have used his data for $D/\lambda \geqslant 0{\cdot}2$, and otherwise Haddock's (1948) cross-sections for 3·0 cm and 18°C. For $\lambda = 1{\cdot}8$ cm we have interpolated between Haddock's data for 0·3 cm and Aden's data.

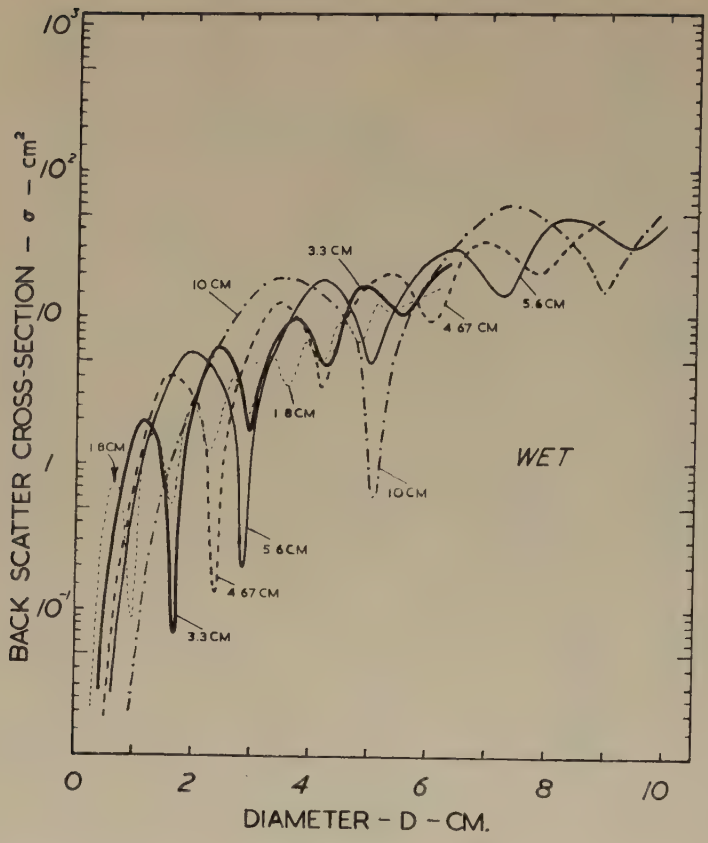

Figure 2 (b). Back-scatter cross-section σ of spheres of liquid water at the indicated wavelengths (used in the text as an approximation to the cross-section of spheres of wet ice).

Atlas *et al.* (1960) found experimentally, and Herman and Battan (1960) confirmed theoretically, that the cross-section of an ice sphere is markedly affected by even a thin, surface-film of liquid water, such as is present during melting and during growth in the 'wet' regime (Ludlam 1958). It appears that, in general, the cross-section is reduced towards that for an all-liquid sphere.

In the experimental work it was found that after melting had set in for a minute or two, the cross-section of a large ice sphere $D > \lambda$ was reduced by about 5 db at 3·3 cm and about 2 db at 4·7 cm, and it was estimated from other experiments that at this stage the thickness of the liquid film was not more than 10^{-2} cm. Frequently the 3·3 cm cross-section was still falling when the measurements were ended, and sometimes had fallen below the all-water value. At first we attributed this to internal melting, but Harper (private communication) points out that his recent laboratory work and the calculations of Herman and Battan indicate that this behaviour is not dependent upon some additional phenomenon. During prolonged melting of an ice sphere in the laboratory, with drop-shedding and a roughly constant thickness of liquid coat, its 3·2 cm cross-section oscillates between values up to a few db above and as much as 13 db below the all-water value. In the range of D/λ from 1 to 2·5 the minima occur at intervals of $D/\lambda \approx 0{\cdot}3$. The oscillations may be due to interference between radiation reflected after being focused on the rear surface of the sphere and that reflected directly at the front surface (from the geometric optics of the interfering rays an interval of $D/\lambda = 0{\cdot}28$ between successive minima can be deduced).

Accordingly, when the water coat is sufficiently thick (but not so thick as to absorb strongly) the cross-section depends mainly upon D/λ. The necessary thickness is about 10^{-2} cm at 3 cm wavelengths, about that found on melting ice particles, and rather more at longer wavelengths. This degree of wetting reduces the 3 cm cross-section of ice spheres of a particular size by as much as 20 db or as little as 5 db; the reflectivity of a group of spheres with diameters varying by a cm or more would probably be decreased to within a few db of that computed from the individual all-water values. At 4·7 cm the observations of Atlas *et al.* (1960) and the calculations of Herman and Battan indicate that the liquid coat thickness must be substantially greater than 10^{-2} cm to produce such pronounced effects (with a thickness of 10^{-2} cm the greatest decreases of cross-section are only to the all-water value – and then only if $D \geqslant 7$ cm).

Evidently the behaviour of hail subject to melting or wet growth is complicated. According to the simple theory of Ludlam (1958) a sufficient amount of melting to provide a coat of thickness 10^{-2} cm occurs while a large stone falls several hundred metres from the melting level, and must be expected to have a large effect (a decrease of 10 db or more) upon the reflectivity at 3 cm, and a slight effect (a decrease of a few db) at considerably longer wavelengths.

The thickness of the liquid film on stones growing in the wet regime is not known, but at least near the 0°C level there may be a ' mushy ' partially-frozen coat a millimetre or more thick (List 1960; Macklin 1960) probably sufficient to bring the back-scatter cross-section close to the all-water value at all wavelengths. We have used this value for all ' wet ' stones, although at the longer wavelengths the film thickness on stones which are melting or barely in the wet growth regime is probably not sufficient to so markedly affect the back-scatter.

4. Reflectivity of hail spectra

For each spectrum of Fig. 1 and for several wavelengths we have computed η, the reflectivity per unit volume ($\mathrm{cm^2\ m^{-3}}$),

$$\eta = \int N_D\, \sigma_D\, dD \quad . \quad . \quad . \quad . \quad . \quad (2)$$

Hailstones of diameter more than 6 cm are too sparse to influence η*.

In the Rayleigh scattering regime we have

$$\eta = 10^{-6}\, \pi^5\, |K|^2\, Z/\lambda^4 \quad . \quad . \quad . \quad . \quad (3)$$

where $K = (m^2 - 1)/(m^2 + 2)$, m being the complex refractive index (for wavelengths from 1·8 to 10 cm and for *liquid* water at temperatures from 0 to 20°C, $|K|^2 = 0{\cdot}93$), and $Z = \Sigma\, ND^6$ ($\mathrm{mm^6\ m^{-3}}$)

When some of the scattering particles are outside the Rayleigh regime it has become customary to express storm reflectivity as Z_e, the equivalent Z, or the value of Z required to satisfy Eq. (3) for the observed value of η; thus

$$Z_e = 3{\cdot}52 \times 10^3\, \lambda^4\, \eta \quad . \quad . \quad . \quad . \quad (4)$$

Using this relation the reflectivity of the hail spectra has been expressed in terms of Z_e and is illustrated in Fig. 3. It appears that whether the particles are wet or dry, concentrations of more than 40 g $\mathrm{m^{-3}}$ are implied by observations of 3 cm intensities of over 10^7 $\mathrm{mm^6\ m^{-3}}$. These concentrations seem so excessive that we made further computations of the reflectivity of some spectra which contain a greater proportion of the more highly reflecting particles, before concluding that the observations point strongly to the local presence of extremely narrow spectra consisting of large dry hail of practically uniform size.

* The publication (Atlas and Ludlam 1960) of limited distribution, copies of which are held in the Society's Library, contains diagrams of cumulative reflectivity showing the progressive increase as the upper limit to the particle size is raised. From these diagrams the reflectivity of any exponential spectrum can readily be interpolated.

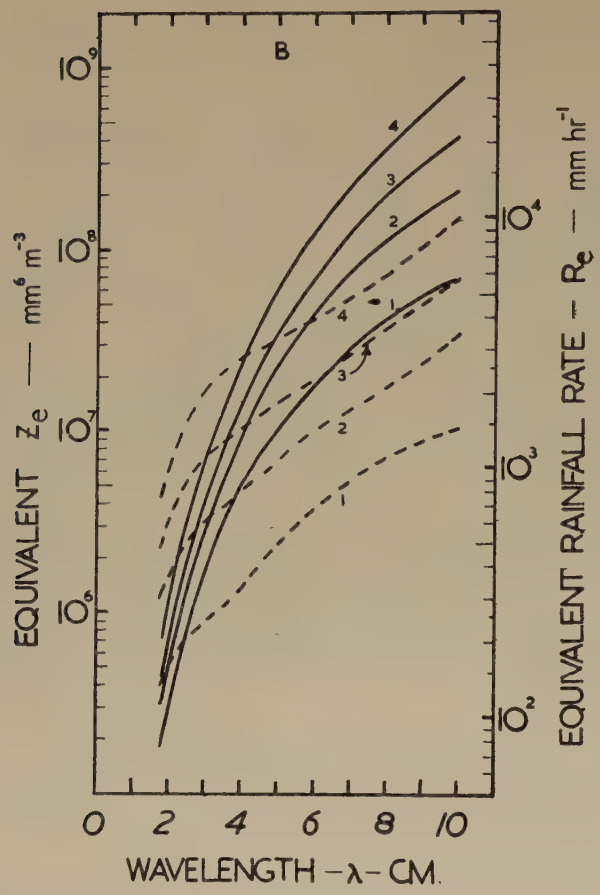

Figure 3. Z_e, the equivalent Z, of the four exponential hail spectra, as a function of wavelength. On the right is a scale of equivalent rainfall rate R_e, derived from the empirical relation ($Z = 486\ R^{1\cdot37}$) of Jones (1956).

Accordingly we have also calculated the reflectivity Z_e of a concentration of 1 g m^{-3} of hail of uniform size and density 0·9 g cm^{-3}, with the results given in Figs. 4 and 5. These show that at a wavelength of about 3 cm, a value of Z_e of 10^7 can be obtained from not much more than 1 g m^{-3} of large dry hailstones; if they are wet, concentrations of at least 10 g m^{-3} are required. The variation of reflectivity with wavelength is so considerable that the possibility arises of using measurements of intensity at several wavelengths to make inferences about the sizes and surface condition of the hailstones present, and we have attempted this with observations made on the Wokingham storm. To interpret the measurements of echo intensity we need to take into account the attenuation due to intervening precipitation, and have therefore calculated that caused by a concentration of 1 g m^{-3} of dry hail of uniform size, as a function of wavelength and stone diameter, with the results shown in Fig. 6*.

5. Echo intensity measurements in the Wokingham storm

This storm has been described by Browning and Ludlam (1960). It was observed at three wavelengths with radars whose characteristics are listed in Table 2. All had been calibrated on the previous day, using a standard target (Atlas and Mossop 1959) and we are confident that the relative measured intensities are accurate to about $\pm 1\frac{1}{2}$ db.

* We used data on the total scattering cross-section provided by Herman and Battan (1961); unfortunately there are as yet no similar data for wet ice spheres.

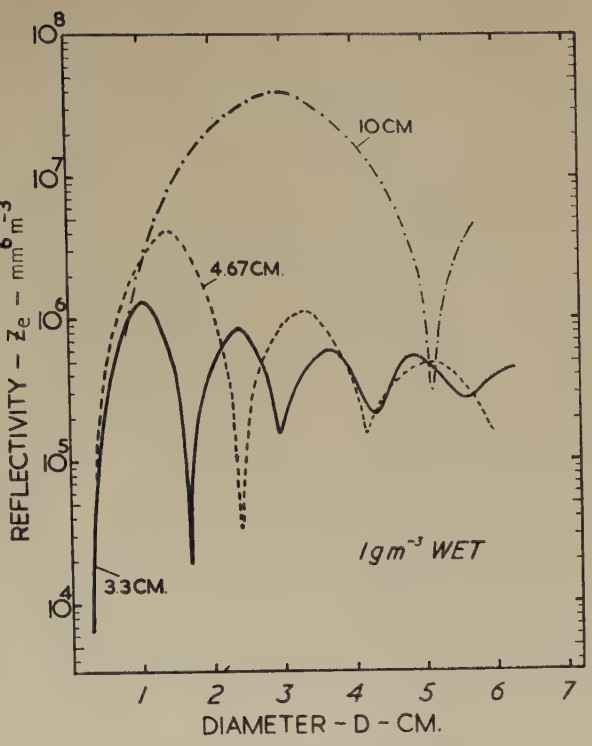

Figure 4. Z_e corresponding to 1 g m^{-3} of wet hailstones of uniform diameter D at three wavelengths.

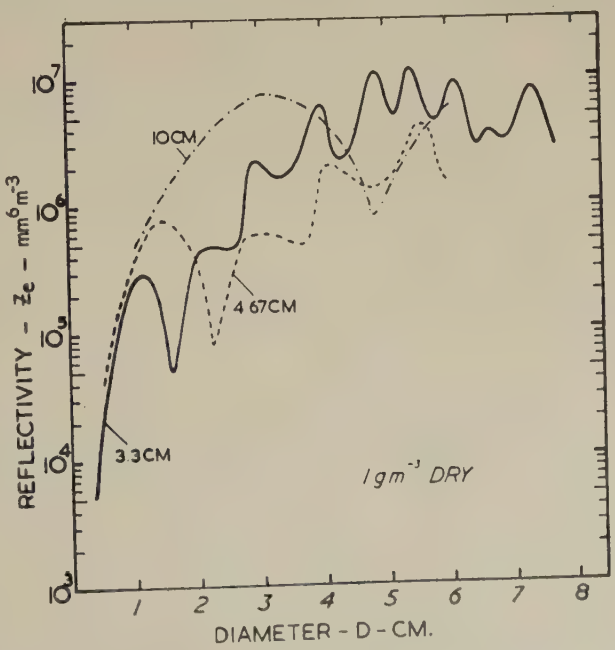

Figure 5. Z_e corresponding to 1 g m^{-3} of dry hailstones of uniform diameter D at three wavelengths.

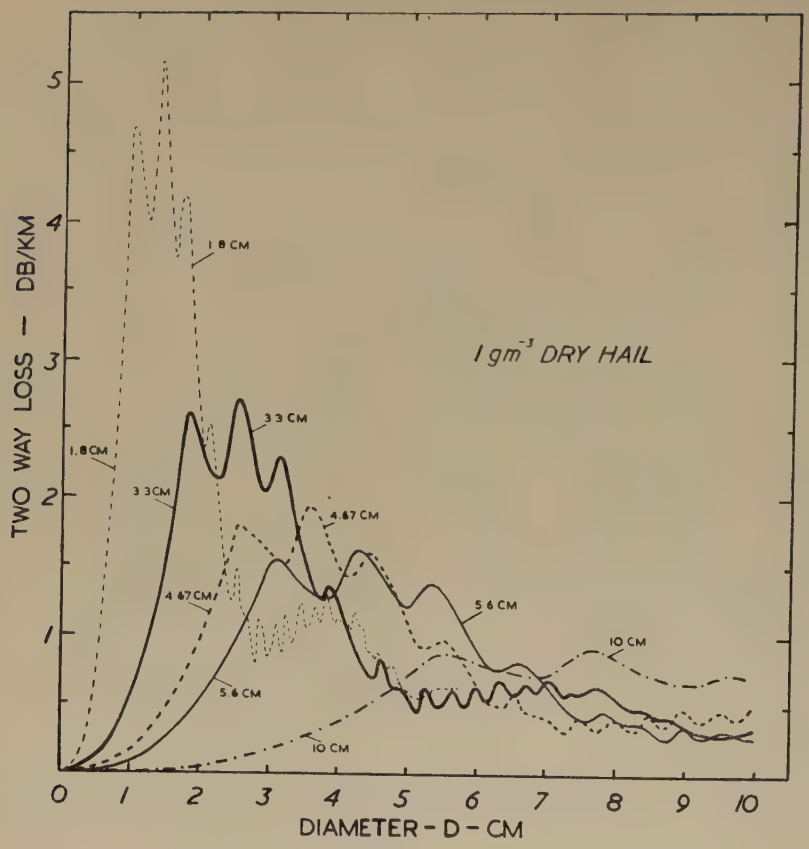

Figure 6. Two-way attenuation, in db/km, at several wavelengths, due to 1 g m^{-3} of dry hailstones of uniform diameter D.

TABLE 2. CHARACTERISTICS OF RADARS USED TO OBSERVE THE WOKINGHAM STORM, 9 JULY 1959

Parameter	TPS-10	MPS-4	Type 13
Wavelength (cm)	3·31	4·67	10
Scan	RHI	RHI, PPI	RHI
Trans. power, peak (kilowatts)	35	140	500*
Min. detectable power (watts)	6×10^{-13}	2×10^{-14}	–
Beam widths (deg.)			
horizontal	2·0	4·9	7·5
vertical	0·7	0·8	1·5
Pulse width (metres)	300	390	600
Polarization	Vert.	Vert.	Vert.
C†	8×10^{-12}	$1{\cdot}2 \times 10^{-11}$	–

* Nominal

† $C = r^2 P_r/Z$, where r is range (mi), P_r is echo power in watts, and Z is reflectivity ($mm^6\ m^{-3}$), from calibration observations on a standard target.

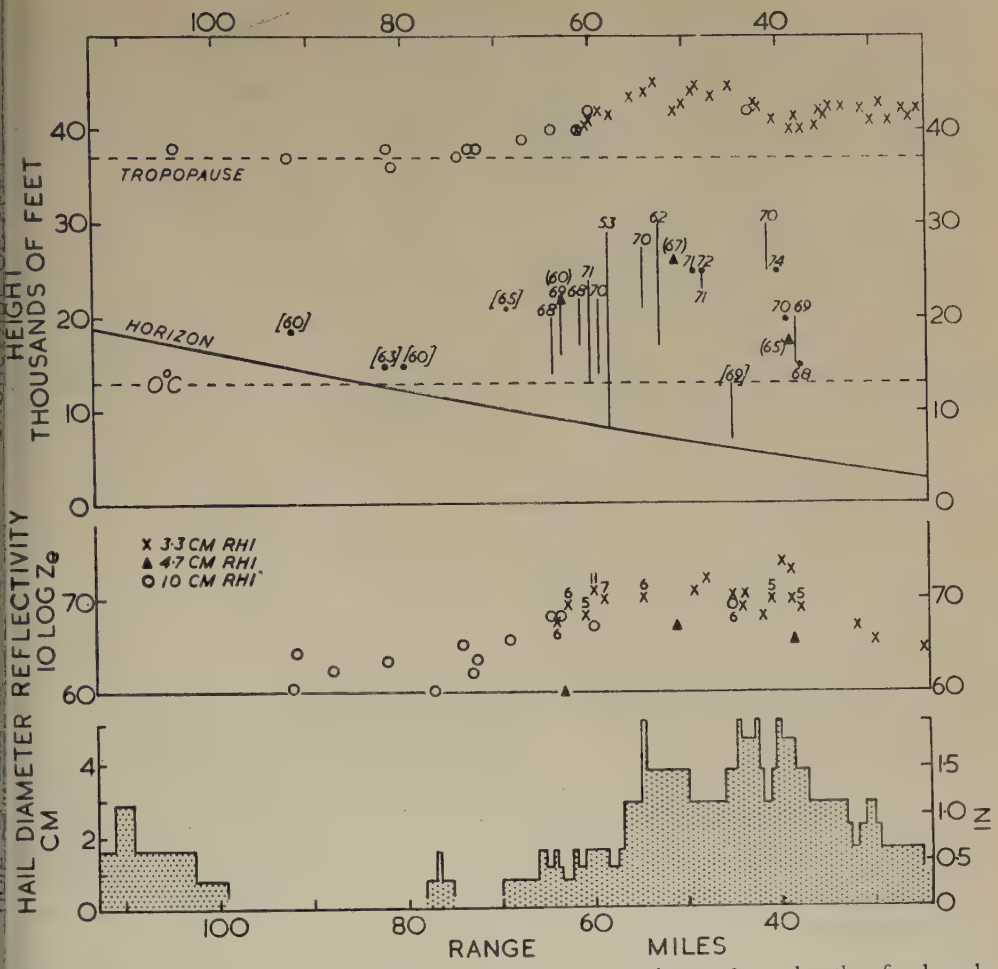

Figure 7. Maximum hail size at the ground (bottom), and observations at 3 wavelengths of radar echoes (above), in the Wokingham storm. In the uppermost diagram the position of the radar horizon, peak echo heights, and the vertical extent of strong echoes are shown (figures beside the last indicate intensity as 10 log Z_e, none or in square or ordinary brackets according to whether the wavelength is 3·3, 10 or 4·7 cm respectively). The middle diagram again shows the intensity of the strongest echoes. Figures beside the crosses indicate the vertical extent, in thousands of feet, of the strongest 3·3 cm echoes.

A summary of the maximum storm reflectivities, the height of the echo tops and the hail size at the ground, as a function of time, is given in Fig. 7*. The 4·7 cm and 10 cm measurements of intensity have been adjusted for incomplete filling of the beam by assuming that the part of the storm containing the strongest echo was 2 mi wide, as indicated by a narrow-beam 10 cm PPI radar. The maximum adjustments were 3 db for the 4·7 cm radar and 6 db for the 10 cm radar, at the range 60 mi, and may have a relative accuracy of $\pm$ 1 db. It is unfortunate that usually only the position and intensity of the strongest echoes, and not the distribution of echo intensity, were observed at 3 cm and 10 cm, and at the latter wavelength were made less frequently. Consequently, comparison of echo intensities can be made only at 1143 and 1203 GMT. The most intense echo lay behind a zone of echo in which Z_e at 4·7 cm was between 10^5 and 10^6 over a distance of about 2 km at 1143 hr and 3 km at 1203 hr. This zone was well above the 0°C level and forward of the storm, and can be attributed to dry hail of any diameters between 0·7 and 3·0 cm in concentrations of about 1 g m^{-3} (Fig. 4), in which the two-way attenuation would amount

* Fig. 7 shows that the onset of hail at the ground is closely associated with increases in the maximum reflectivity at all levels and in the height of the echo tops. The period of hail of diameter more than 3 cm matches closely that in which the 3·3 cm Z_e exceeds 10^7.

to about 1 db/km at 4·7 cm and 1½ db/km at 3·3 cm (Fig. 6). At 10 cm the total attenuati is only ½ db or less. We cannot be very confident about the implied corrections to t maximum values of Z_e (3 and 4½ db respectively at 3·3 cm, and 2 and 3 db respectively 4·7 cm), but it is unlikely that they could be more than twice as great. The adjusted valu are listed in Table 3; the errors in the basic measurements, added to those in the adjustmer for beam-width effects and attenuation, are unlikely to cause relative errors exceedi 4 db. Bearing this in mind, we can state :

(a) at 1143 hr the most intense echo is a little stronger at 3·3 cm than at 10 c and considerably stronger than at 4·7 cm and (b) at 1203 hr it is about 4 db stronger 3·3 cm than at either 10 cm or 4·7 cm.

TABLE 3. MAXIMUM ECHO INTENSITY, EXPRESSED AS 10 log Z_e ($mm^6\ m^{-3}$), IN WOKINGHAM STORM AT VARI WAVELENGTHS, AFTER CORRECTION FOR INCOMPLETE FILLING OF THE BEAM (A), AND ALSO FOR ATTENUATION

(GMT)	Wavelength :	3·3	4·67	10 cm
1143	A	68	60	67
	B	71	62	67
1203	A	71	67½	69½
	B	75½	70½	70

These features are not reconcilable with the computed intensities of the exponenti spectra of hail, which increase markedly with wavelength (Fig. 3). However, referri to Figs 4 and 5 we see that statements (a) and (b) can be satisfied if the hail is dry and uniform diameter about 3·9 and 5 cm respectively. Considering the fallspeeds of spheric hailstones of these sizes (averaging 30 m sec^{-1}, but effectively reduced to 20 m sec^{-1} whi traversing the updraught) and the forward speed of the storm, we should expect them reach the ground about 3½ mi nearer the radar than the positions of the associated echo (and therefore at 57½ mi and 45½ mi respectively). The ground observations indicate th stones of just these diameters are first encountered at ranges of 55 mi and 44 m respectively*.

Accepting that the most intense echoes were due to hailstones of uniform size, th implied concentrations are about 1½ g m^{-3} and 5 g m^{-3}. These might still be regarded a rather high, but they seem much more reasonable than those of more than 40 g m associated with exponential spectra. The implied number concentrations are abou 6×10^{-2} m^{-3} and $7{\cdot}5 \times 10^{-2}$ m^{-3} respectively, which may be compared with the estimate made by Ludlam and Macklin for the highest concentrations occurring near the groun namely about 2×10^{-2} and 6×10^{-4} m^{-3} (Fig. 1). The agreement in the first case i reasonable, but in the second the inferred value is about two orders of magnitude greate than the highest they obtained; we are not able to give any satisfactory explanation fo this discrepancy. This and the evidence that the stones may have descended more slowl than assumed, lead us to mention that the largest stones collected from the storm ha the form of spheroids rather than spheres. It is known (Harper, private communicatio that this could have the effect of somewhat reducing the concentration of stones require to provide the echo intensity observed on the vertically-polarised radars, besides reducin the fallspeeds. Unfortunately relevant data are too sparse yet to be useful. The dis crepancies mentioned throw some doubt on our interpretation of the intense echoes, bu the potential value of measurements at several wavelengths is clear.

The deduction that the most intense echoes were returned by the largest hail, presen in almost uniform size and in a dry condition, is supported by the dynamical model o the storm developed by Ludlam and Browning (1960). In this the base of the updraugh is located over the squall front, some miles ahead of the region of strong updraught a

* The discrepancy of about 2 miles, in each case, is significant and implies that the mean fallspeeds were only about 70 per cen of those appropriate to smooth spheres of the specified diameters.

dle levels and the position of the most intense echo; at still higher levels, near the opause, the updraught diminishes and the air spreads forward of the storm. Hailstones released from these levels and fall clear of the updraught, evaporation drying their faces. The wind decreases markedly towards the ground, and in consequence of their erent fallspeeds the stones tend to become sorted according to size, especially the gest which reach middle levels close to the updraught. Smaller stones arrive at these els only at increasingly great distances, and probably in broader spectra since they y leave the updraught over a greater height range. The smallest may evaporate com-tely, others reach the ground as rain ahead of and to the left of the storm, while the ger stones re-enter the updraught in the lower middle levels.

This re-entry can explain the rapid decrease of echo intensity below the strongest o, which is difficult to account for in any other way, as an effect of wetting. Even some tance above the 0°C level the largest hail can be expected to enter the wet growth ime, whereupon its reflectivity can be anticipated to decrease by an order of magnitude 3 cm and by several db at 4·7 cm. Its contribution to Z_e should then diminish to about ^{6} mm^6 m^{-3} or less at both wavelengths. The observed maximum intensities near the C level were indeed about or just in excess of 10^6 mm^6 m^{-3}. It is difficult to discuss m without the data to take into account the effect of attenuation, but it is clear that the no from large wet hail would be only comparable with that from the intense nearby fall small wet hail and rain. From the total rainfall ($\frac{1}{2}$ to 1 in.) recorded at places most verely affected and the known duration (several minutes) the maximum precipitation te can be estimated as about 200 mm hr^{-1}, corresponding to a concentration of rain ar the ground of about 7 g m^{-3}. This, according to the empirical relation of Jones (1956) tween echo intensity Z and rainfall rate R ($Z = 486\,R^{1\cdot37}$), would produce an echo tensity of about 7×10^5 mm^6 m^{-3}. According to Fig. 5, only at particular diameters se to 1 and 2·3 cm (at a wavelength of 3·3 cm), or 1·4 and 3·3 cm (at 4·7 cm), could similar concentration of wet hail produce echo intensities substantially greater than ^{6} mm^6 m^{-3}.

It is interesting that the echo intensity varied only slightly below about 20,000 ft at e 10 cm wavelength, for which the hail may always appear to be dry, and that the tensity maximum tended to be in the range 10,000 to 15,000 ft, not far from the 0°C vel. Its value of nearly 10^7 mm^6 m^{-3} is not incompatible with the exponential hail ectrum containing a concentration of a few g m^{-3}. Since this value exceeds that corres-nding to rainfall at the world record rate of 1,000 mm hr^{-1}, it implies the presence of mall hail, but cannot indicate large hail, which in any reasonable concentration makes nly a small contribution to the echo (unless its diameter is close to 3 cm).

6. General characteristics of hailstorm echoes

Ludlam and Browning (1960) suggest that their dynamical model of the Wokingham torm is representative of middle-latitude severe storms, and in particular that the production f large hail demands an updraught having a configuration such that small hail can re-enter after falling from a higher level. It may then always be possible for the large hail to ecome dry and separated from the other precipitation before itself re-entering the up-raught and becoming wet and less highly reflective. Consequently, the distribution of cho intensity in the Wokingham storm, which at 3 cm is similar to that observed re-eatedly in severe storms by Donaldson, is likely to be typical of storms producing large hail.

According to the model mentioned the maximum size of hailstones is determined y the maximum updraught speed reached in the zone of super-cooling. This is probably ot closely related to the (low-level) flux of water vapour into the storm, which limits the otal precipitation rate. It can therefore be expected that not all severe storms produce arge hail, and that in general the low-level 3 cm Z_e will be limited to about 10^6 mm^6 m^{-3}. When large hail is formed and can become dry aloft even in a modest concentration, it may be recognisable at 3 cm by an abnormally strong echo. Some hail at the ground is to be expected when its intensity considerably exceeds 10^6 mm^6 m^{-3} (implying stones of

diameter 2 cm or more aloft which can melt only partially before reaching the ground large hail is to be expected when the intensity exceeds 10^7 mm^6 m^{-3} (diameter aloft 4 cm more, negligibly reduced by melting during fall). The echo intensity at 4·7 cm is a rather less sensitive indicator of large hail, and at 10 cm high intensities are suggestive only of the presence of small hail.

Acknowledgments

We are pleased to thank the Director-General of the Meteorological Office for the provision of staff and facilities at the East Hill Radar Station, from which our observations were made. The research on severe storms of which the work is a part, is supported by the Geophysics Research Directorate, Air Force Cambridge Research Center of the Air Research and Development Command, United States Air Force. The authors are indebted to the following personnel who participated in the observational programme: Messrs. W. G. Harper (Officer-in-Charge), N. Coupe, A. Denham, and J. Beimers of the East Hill Radar Station; Dr. and Mrs. S. C. Mossop of the South African Council for Scientific and Industrial Research.

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551.508.765 : 551.574.12

Fog-droplet sampling using a modified impactor technique

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(Manuscript received 1 February 1961; communicated by B. J. Mason)

Summary

Possible errors in sampling fog droplets with the cascade impactor are discussed and data are presented hich were obtained from 25 fogs by the use of a specially modified impactor in a small portable wind-tunnel. very wide range of drop diameter (< 1 to $> 100\ \mu$) was found, with great numbers of very small drops d great variability in size-distribution. Agreement between the visual range as calculated from the drop easurements and the actual visual range was usually closer than a factor of two, though in some cases there as no such agreement.

1. Introduction

A criterion for the completeness and accuracy of a sample of fog or cloud, when e opacity is due to water droplets only, is that it should yield values of drop concentration, ze distribution and water content, from which the computed value of visual range is n agreement with direct observation, with no assumptions other than the correctness of isibility theory. Mason (1957) considers that direct methods of sampling, involving the ounting and sizing of individual drops from a known volume of fog or cloud, should e the most accurate. No published work on fogs, however, appears to meet the above riterion fully, because of the difficulty in obtaining a true sample in direct methods and ecause indirect (e.g., radiation transmission) methods involve various assumptions.

The 'cascade impactor' (May 1945) was specifically designed for sampling windborne droplets in the size range of roughly 100 to $\frac{1}{2}\ \mu$ diameter, and has been widely used in og work because of its apparent suitability. However, results with it are suspect because Elton and his co-workers (1956), Liddell and Wootten (1956), Stewart (1957), Todsen (1959) and Pedersen and Todsen (1960) all found that attenuation coefficients or liquid-water contents of fogs as indicated by the cascade impactor were usually lower by one or even two orders of magnitude than was expected.

The present paper discusses the possible sampling deficiencies of the cascade impactor and their elimination by various modifications. The drop-size distribution data of fogs so obtained are probably more complete than any so far reported and they are often consistent with simultaneous measurements of visual range and water content.

2. Use of the cascade impactor with water droplets

The cascade impactor as manufactured by Casella & Co., Fitzroy Square, London, s an instrument through which particle, or droplet-laden air, is sucked by means of a pump or compressed air ejector. In its passage the air passes in series through four successively finer slit-shaped jets which throw out size-graded fractions of the aerosol on to four microscope slides. The larger droplets are impacted at relatively low velocities so that shatter of drops has never been observed. The size grading also minimizes droplet overlap and assists microscopic analysis.

The characteristics of the four stages of the Casella impactor when operated at 17·5 litres/min are shown in Table 1.

Column 4 in Table 1 presents the effective limits of drop diameters found on each stage and the central figure is the size at which 50 per cent of the drops are impacted, the remainder being caught on the next stage. The actual droplet retention curves for each stage are known with considerable precision and may be found in Casella's instruction pamphlet for the instrument.

TABLE 1. CASELLA IMPACTOR CHARACTERISTICS

Stage	Slit dimensions (mm)	Jet velocity (m sec^{-1})	Range of unit-density droplets impacted (μ)	Pressure drop across jet (mb)
1	19 × 6·5	2·2	Aerosol max size — 12 — 8	0·06
2	14 × 2	10·2	19 — 4 — 2	1·1
3	14 × 0·75	27·5	7 — 1·5 — 0·8	7·1
4	14 × 0·27	77	2 — 0·5 — 0·3	75

The most convenient field methods for obtaining permanent water-droplet sampl on glass slides are :

(*a*) *Magnesium-oxide method*

The preparation and use of slides coated with magnesium-oxide smoke has bee fully described by May (1950) and various practical applications in cloud studies a given by Mason (1957).

In the cascade impactor, the method can be used down to about 4 μ on stage 3 provide that (i) the oxide layer is cleaned well off the edges of the slides so that it is not rubbe off by the slide-locating runners, (ii) all loose oxide is tapped off the slide before insertio (iii) on stages 2 and 3 the layer is no thicker than about 30 μ. Magnesium oxide canno be used on stage 4 as the fast air-jet destroys the layer.

(*b*) *Stain method*

Liddell and Wootten (1957) describe a method in which a thin film of a solutio of gelatin plus a purified green dye is spread on a glass slide and allowed to dry. Whe water droplets strike this dry film, a permanent darker ring is left proportional to the drople size, before evaporating. When well prepared the film can detect droplets down to 1 diameter, is therefore more sensitive than the magnesium-oxide method and is ver convenient for field use.

The stain method is greatly improved in simplicity and sensitivity by eliminatin the rather troublesome dye and using only a pure gelatin film, the small stains bein viewed by phase-contrast (May 1959). There is excellent definition of the larger drople stains in the normal light microscope but the phase-contrast microscope reveals th smaller droplet traces (< 10 μ) very clearly down to stains of 1-1$\frac{1}{2}$ μ diameter, corres ponding to droplets of 0·6-1 μ. The disadvantage of the method is that the ratio of drop size/stain-size, although constant for a given slide, is not necessarily constant from sampl to sample and values between 0·55 and 0·65 have been found. The variation may be du to the effects of surface tension, humidity, temperature, etc. The calibration can be checke by having a small magnesium-oxide portion on each slide and this is now the metho of choice for any slide exposed to fog, whether in an impactor or not. The excellen optical quality of the gelatin film enables any solid residue left by the droplets on drying to be clearly seen (e.g., condensation nuclei). Under a cover-slip the sample is permanent

3. SAMPLING ERRORS OF THE CASCADE IMPACTOR AND THEIR AVOIDANCE

(*a*) *Large-droplet losses*

Ideally all droplets originally contained in the volume of air sucked into the impactor would be deposited on the slides but, in practice, the high inertia of the larger droplets makes them very prone to be lost by deviating from the streamlines entering the orifice or by striking the internal walls. The intake orifice was designed to give minimum inertial loss of droplets when facing directly into a 3·5 m sec^{-1} wind and sucking at 17·5 litres/min.

Wind-tunnel studies (May 1956) confirmed that intake efficiency, in the non-turbulent wind-tunnel air, was close to 100 per cent for droplets at least up to 50 μ diameter in 3·5 to 4·5 m sec^{-1} winds. In quite stagnant air the intake efficiency was similarly high for all droplets tested. The efficiency could fall to around 50 per cent for 50 μ drops in 1 to 2 m sec^{-1} winds and would be worse for larger drops. Above 4·5 m sec^{-1} the efficiency exceeded 100 per cent for the larger drops. The ideal conditions readily produced in the laboratory of laminar flow or quiet air never occur in the field. It is true that there is often an absence of wind in fog, but an observation of the droplets in a lamp beam in calm night-fog will nearly always show them to be in a state of rapid random swirling motion. Such motion is the opposite of the conditions required to obtain a good sample of the larger droplets which are predominant in many fogs. Steady winds of 3·5 to 4·5 m sec^{-1} are rare and it seems, therefore, that to ensure a good sample, an artificial wind of the optimum speed must be provided. This can readily be done by means of the small portable wind tunnel described later, which should enable droplets up to at least 100 μ to be sampled with small error.

(*b*) *Small-droplet losses*

Very small changes in relative humidity can cause small water droplets to evaporate or grow very rapidly. Such changes are possible in the Casella impactor, bearing in mind that the time a droplet takes to reach the last slide is about 0·25 sec during which the pressure finally falls by 7 mb (Table 1, Col. 5), while the droplets are subjected to severe wind shear and turbulence in metal tubes which may not be exactly at ambient temperature. Practical experience in fogs (e.g., Stewart 1957) indicates that small droplets do, in fact, evaporate after stage 2 and are lost. Therefore, in the present work, an impactor with only two stages was used, namely, the normal stages 1 and 3 of Table 1. The passage time to the new second stage is now about 0·125 sec in a pressure drop of only 0·06 mb so that droplets at least down to 1 μ are unlikely to change in size provided that the instrument is allowed adequate time to reach ambient temperature before use. From Table 1 we see that the new second stage will catch droplets from 19 μ to 0·8 μ but retains only 50 per cent of 1·5 μ droplets. By applying corrections from the known curves of droplet retention vs diameter for this stage we can estimate the numbers of droplets occurring down to a little below 1 μ which is the lower limit of the present technique. Stage 4, or a millipore filter, can be added if it is desired to collect solid atmospheric pollution.

(*c*) *Sampling-time errors*

To be countable, a droplet sample on a slide must have the droplets or their traces well spaced-out with infrequent overlaps. In relation to the droplet concentration in the thicker fogs the normal sampling rate of the impactor (17·5 litres/min) is high, so that swamping of a slide by an excessive sample easily occurs unless an extremely short sampling period, sometimes less than 1 sec is employed. Such short sampling periods are difficult to obtain and measure and give starting, stopping and dead-space errors. Also, snap samples are of doubtful value as it is unlikely that fogs are homogeneous in concentration. These difficulties may be completely removed by giving the sampling slides a uniform-velocity reciprocating motion of short period, across the line of the slit-jet. The sample is thus spread out over a large area of the slide so that the slide capacity and the sampling time are greatly increased. When the sampling time embraces several cycles of reciprocation any one part of the impact area can be regarded as an integrated and representative sample of the whole of the sampling period.

To sum up, a two-stage reciprocating-slide impactor with gelatin-phase-contrast drop detection, mounted in a small portable wind tunnel should eliminate inherent errors of the standard 4-stage impactor and should be capable of giving fog samples of good accuracy down to 1 μ diameter.

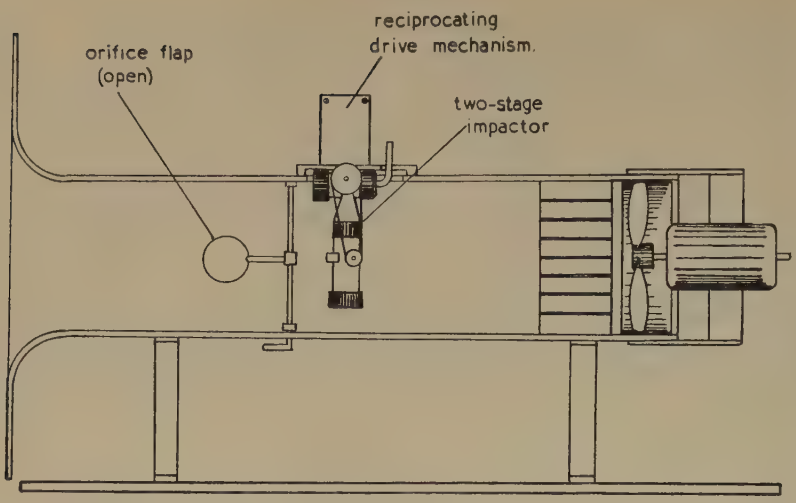

Figure 1. Portable wind tunnel in longitudinal section.

4. Description of apparatus used

(*a*) *The portable wind-tunnel*

A diagram of this is shown in Fig. 1. The cross-section of the parallel portion is 20 cm square. To give a reasonable laminar air-flow, the mouth is flared out smoothly to form the 60 cm square plate as shown. This large plate is intended to act as a damper on unwanted natural air movement and turbulence in the vicinity of the tunnel mouth, especially when sampling in a wind. To facilitate internal adjustments etc., one side of the tunnel is made of perspex. A honeycomb in front of the fan is intended to prevent the possibility of the build up of swirl in front of the fan. The fan is driven by a mains-operated fractional horsepower motor controlled by a rheostat and gives a maximum airspeed of 5 m sec^{-1}.

The ' orifice flap ' shown in Fig. 1 is a metal disc on the end of an arm pivoted at the side of the tunnel. An external lever enables the flap to be moved from its first position of covering the sampling orifice to its second position out of the way, flush along the tunnel wall as in Fig. 1. Thus the sampling orifice can be exposed to the fog during the defined sampling period only, an essential precaution as the larger fog droplets are otherwise projected into the orifice by the air stream when sampling is not taking place.

The tunnel is operated facing into the wind, if any, with frequent checks on internal air speed by a velometer or the very simple Hooke type of swinging-vane anemometer (May 1943).

An obvious requirement when sampling inside a wind-tunnel is that the tunnel itself should have a high efficiency of droplet intake. This should be so, provided that the tunnel bore is large in relation to the ' stopping ' or ' projection ' distance of the largest droplets from the speeds at which they approach the tunnel mouth. The stopping distance for a 100 μ droplet at 4 m sec^{-1} is 11 cm for example. Practical considerations have limited the existing tunnel to 20 cm square section and it is possible that this is not really large enough to avoid error in the intake of the larger droplets.

Sampling orifices in the tunnel are placed in the centre of the cross-section, about 40 cm from the tunnel mouth. This is far enough for droplets to pick up the full stream velocity and for a damping effect on the normal undesirable ambient air turbulence to be exerted, but not so far that the point is in the droplet ' shadow ' under the tunnel roof caused by sedimentation effects. A 150 μ droplet, for example, would fall about 6·5 cm per 40 cm of travel through the tunnel at 4 m sec^{-1}. A hole in the roof enables sampling apparatus to be set up in the correct position, as in Fig. 1.

The tunnel described is very light in weight and is easily transported on a small boratory trolley, a lower tray of which can hold the necessary batteries and pumps.

(*b*) *The reciprocating-slide impactor*

This is an adaptation of the ' moving-slide impactor ' (May 1956) in which a mechan- m moves the microscope slides continuously across the jets. In the present instrument motor-driven cylindrical cam imparts a uniform velocity reciprocating motion to each de with a period of about 2 sec and an amplitude of 4 cm for the first slide and 2 cm r the second slide.

One second of sampling for every ten yards of visibility gives a convenient droplet ensity on the slides for counting. Constructional details of the mechanism are available n request, but a greatly simplified sampler is being developed which is a two-stage, nd operated, continuously rotating drum impactor.

When counting and sizing the droplet traces, the extreme ends of the deposit are oided because there are too many droplets where backlash halts the slides momentarily. he first jet has a tendency to deposit more droplets at the sides of the microscope slide an in the centre. Therefore counted areas are strips right across the slide at right-angles the direction of motion. Counting and sizing is done with the aid of the eyepiece raticule designed by the writer for this purpose (May 1945). Counted strip widths are elineated by the lines in the graticule and it is usual to choose the width or number of rips so that about 300 droplets are counted per slide. Errors in counting the largest roplets usually have a big effect on the final result, and accuracy is increased by counting s big an area of the first slide as possible by a second scan at low magnification, dealing nly with the largest droplets. On the second slide, to overcome droplet density variations ue to imperfections in the reciprocating mechanism, strips are counted along the direction f travel for as great a length as can be done.

Counting is usually done with a $\times$ 20 objective for the first stage and $\times$ 40 or $\times$ 95 or the last stage, adjusting the microscope magnification so that overlapping size ranges re the same. Knowing the area counted and the total area of the deposit, the total number f droplets in each size range on each slide are computed from the completed counts with corrections from the retention curve for the $< 2\ \mu$ droplets) and summed to give he total distribution. Conventional numerical methods are applied to the total distribu- ion to give the required parameters, in the manner described for example by Dallavalle 1947).

5. Other methods of direct fog sampling

(*a*) *For drop-size distribution*

As a check on impactor sampling, a method which at first sight seems attractively imple and free from errors is to set up in the open fog a whirling arm on the end of which s a narrow glass ribbon or slide for collecting the droplets. The glass can be treated for detecting the droplets as already described. Alternatively, the ribbons can be exposed to he moving air of a wind tunnel by projection from a sheath through the tunnel wall. According to the widely accepted figures of Langmuir and Blodgett (1946) a 1 mm wide ribbon moving through air at 4·5 m sec^{-1} should collect nearly all droplets above about 7 μ and give useful information down to about 3 μ. In practice, both methods of exposure were found to give, in relation to the impactor, extremely low samples seriously deficient in small droplets, and the technique was abandoned. It seemed that the predicted Langmuir and Blodgett collection efficiencies for ribbons were too high, a point which is now being investigated in the laboratory, with similar results. Also the small area available for impact collected a very inconveniently small sample. Dessens (1946) took this principle to its limits by collecting on spider webs but his method seems to have its principal use in sampling haze particles, etc. The method of simply waving a treated microscope slide in the fog as has been done in several earlier investigations is, of course, useless for collecting the smaller drops, and gives no information on drop concentration in the air.

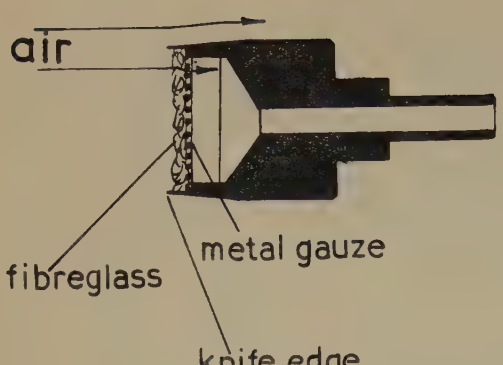

Figure 2. Isokinetic fibreglass filter.

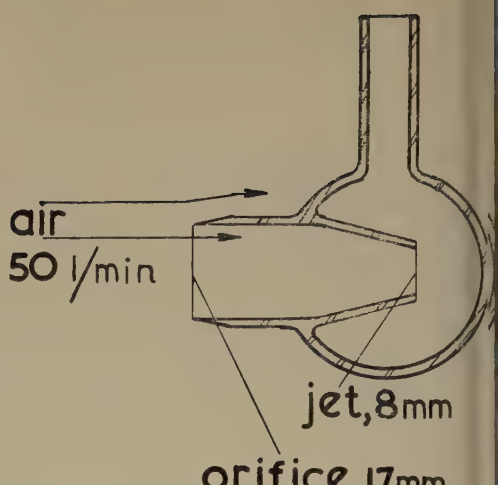

Figure 3. Glass impactor.

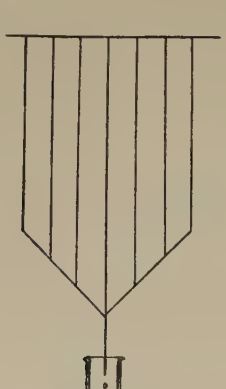

Figure 4. Droplet-collecting grid (diagrammatic).

(b) To measure total water content

A reliable measure of this quantity would give a valuable check on the impactor recoveries, so efforts have been made to devise a convenient physical method for collecting and weighing fog water. Technically, the guard-ring mounted wire screen collector of Radford (1938) seems to be a perfect apparatus for the larger fog droplets in which most of the mass of water is usually found. It was thought to be unsuitable in the present case, however, as it requires a long sampling period which would bear no useful relation to the short-period impactor samples taken in the rapidly varying local fogs. In the portable wind tunnel it seems logical to use a knife-edged isokinetic orifice with a suitable high efficiency non-hygroscopic filter on which small water deposits can be weighed on a chemical balance. Such a device should give an absolute sample provided that there is no mechanical or evaporative loss of water from the filter. A diagram of the arrangement used is shown in Fig. 2. The filter is a thin pad of fine fibre-glass which has a 100 per cent filtration efficiency down to below 1 μ. The pad is supported at the extreme front of the tube by a wire mesh base so that there are no intake tube errors. The knife-edged tube itself is used as the tool for cutting out the pads, thus giving a fit free from edge leaks. Glass wool adsorbs water vapour, therefore the pads are of minimal thickness and the weighing technique is to weight the pads plus fog water in a closed vessel soon after sampling, then to dry off completely in a desiccator followed by a period in a water-saturated atmosphere before

losing the vessel again for the tare weight. In this way the pad is in the same state of ater-vapour adsorption with and without fog water. Too much water must not be collected n the pad or the water masses may be drawn right through by the air flow. In practice, p to 20 mgm cm^{-2} of filter surface has been safe. In the unsaturated air of clearing fogs here is evaporative loss from the pads and the method is obviously only suitable for stable onditions in non-hygroscopic fogs. Even then the possibility remains of the forced entilation past the collected water causing some evaporation. This is one of the many oints which have not yet been fully studied owing to the transient nature and rarity f the local fogs.

Fog filters of 19 mm diameter, operating at 50 litres/min and of 49 mm diameter t 310 litres/min both isokinetic at 2·7 m sec^{-1}, have been used. The larger size seems referable because of the heavier sample.

As most fogs seem to have nearly all of their suspended water in droplets larger han 5 μ, a high flow-rate impactor capable of collecting down to this diameter only and herefore of quite low pressure and evaporation loss, should give reliable samples. Such device made from glass is shown in Fig. 3. Air is sucked through at 50 litres/min and vater in bulk collects in the bulb and in the intake tube. After sampling, the intake and outlet holes are sealed with bungs, the outside wiped dry and the whole weighed. A dry are weight is subsequently made.

A simple arrangement for longer-period collection in stable fogs is to arrange a grid of vertical 1 mm diameter wires across the tunnel. This, in a 4·5 m sec^{-1} wind, will ntercept nearly all the droplets down to about 15 μ diameter which would pass through he projected area of the wire. Usually less than 10 per cent of the total water is in droplets below 15 μ, so the collection of the wires should give a figure close to the real value, in he absence of evaporation. In a practical application of this system 21 vertical 1 mm wires, 1 cm apart, had their lower ends connected to a common point in the manner ndicated in Fig. 4, which for clarity, has a reduced number of wires. From the common point another wire leads down into a small water-collecting weighing bottle held in a hole in the floor of the tunnel. In practice, the grid slopes downwards and backwards at 45° so that impacted water is helped on its downward passage to the bottle by the wind. The wires must of course be clean and grease-free and are soaped before use to aid water drainage. To take a sample, the grid is set up in the tunnel until water starts dripping off the end and the drips are then collected for a known time, usually 5-10 min, when the order of a gram of liquid will be collected. In principle, this wire grid is the same as Radford's apparatus but has a more sensitive water-weighing arrangement.

6. Results

A summary of the results obtained to date is given in Table 2. All samples were taken from the same position at the side of a building on open Salisbury Plain downland. Notes on column headings are :

Col. 3. Visual range. Estimated by reference to a row of thick lamp-posts running from the sampling point. No instrumental measurements were taken.

Col. 4. Drop-size range. Extremes of the sizes found in the impactor samples; usually the upper limit was also checked by a waved microscope slide in the fog.

Col. 5. Number median diameter. The 50 per cent intercept on a smoothed cumulative number distribution curve on logarithmic probability paper.

Col. 6. Volume median diameter. As for Col. 5, but from the cumulative volume distribution curve.

Col. 7. Mean area diameter, d_a. Value of $(\Sigma nd^2/\Sigma n)^{\frac{1}{2}}$, so that if N is the total number of drops per unit volume, $\pi Nd_a^2/4$ is the total cross-sectional area of the fog water in that volume.

Col. 8. Mean volume diameter, d_v. Value of $(\Sigma nd^3/\Sigma n)^{\frac{1}{3}}$, so that $\pi Nd_v^3/6$ is thi concentration of water in the air.

Col. 9. Number of drops per cm³ (N). Summation of all the drops on the impacto slides as computed from the number counted, divided by the volume of air sampled

Col. 10. Water content from impactor. Obtained as described for Col. 8.

Col. 11. Water content by weighing. Figures obtained by methods previously describe as follows : † fibre-glass fog filter 50 l/min; ′ fibre-glass fog filter 300 l/min; * glas impactor; ″ grid of fine wires.

Col. 12. Calculated σ_D. If a beam of light of initial flux I_0 passes through an attenuat ing medium for a distance x the final flux I is given by $I = I_0 \exp(-x\sigma)$. The extinctio coefficient σ is due to scattering and absorption. In the case of clean droplets, which have negligible absorption σ_D = projected area of drops per unit volume × the ' scattering area ratio,' K. $K = 2$ for drops larger than about 2 μ diameter (Middleton 1952). Hence $\sigma_D = \pi Nd_a^2/2$. A factor of $\times 10^5$ brings this to the usual unit of Km^{-1}, with d_a in cm.

Col. 13. Observed σ. Derived from the so-called Koschmieder's formula, $\sigma = 3{\cdot}9/v$ where v is visual range in kilometres.

Cols. 14-16. Best's equation. n and a are the constants in the equation $1 - F = \exp[-(x/a)^n]$ (Best 1951) where F is the fraction of liquid water as drops smaller than x microns.

Table 2 covers the period of evolution of the sampling technique described in this paper. All samples were taken in the portable wind tunnel, but in the first three fogs, a stage 1 and stage 3 impactor with continuously moving slides (May 1956), and with slides coated with the green dye/gelatin detector (Liddell and Wootten 1957) was used. From 4 December 1957 the reciprocating slide impactor has been in use, and the gelatin-phase-contrast slides date from 25 November 1958. From 25 January 1960 a stage 4 impactor has been added to the reciprocating slide instrument but only on 8 September 1960 did this detect droplets.

Most of the non-persistent fogs were of the radiation type, formed overnight and sampled at about 1000 hr when they were about to disperse or actually dispersing.

7. Discussion of results

(a) *Drop concentration and size distribution*

The very wide range of drop diameters given in Col. 4, Table 2 will be noted (320 μ-0·5 μ). No upper limit to a fog droplet can be defined, as the spectrum can merge continuously into that of drizzle or rain when these accompany the fog. At the lower end of the scale, very large numbers of drops smaller than 2 μ diameter were found. In published work to date, only Liddell and Wootten (1957), Todsen (1959) and Pedersen and Todsen (1960), who all used Casella impactors, appear to have found large numbers of small drops in fog samples, though Johnson, Eldridge and Terrell (1954) inferred large numbers of small drops in mountain clouds from infra-red transmissometer measurements. The highest number of drops per cm³ smaller than 2 μ in the present work was 870 on 29 January 1959, and in many of the other fogs numbers running into hundreds were found. However, there are fogs, notably 25 January 1960 and 4 September 1958, where there were no small drops (note the number median diameters in Col. 5) and the overall picture is that fogs have great variability in drop size and size-distribution. The volume-distribution curves of Fig. 5 also demonstrate this wide variability.

A minority of the drop distributions were bimodal with peaks in the ranges 1-2 μ and 15-25 μ and a trough between 6 and 10 μ. Bimodal distributions were also found and discussed by Pedersen and Todsen (1960).

The numbers of drops per unit volume in Col. 9 do not correlate with visual range because of the wide variability in size distribution. The high value of some of these numbers is entirely due to the predominance of the small drops revealed by the sensitive sampling method.

Best (1951) has suggested the equation $1 - F = \exp[-(x/a)^n]$ for fog and cloud drop-size distributions. F is the fraction of the volume of water in the air comprised by drops with diameter less than x, and a and n are constants. The figures in Cols. 14-16 were computed to test this equation in the present work. For each fog, $\log \log_{10}(1/1 - F)$ was plotted against $\log_{10} x$ as in Fig. 5. Where it seemed reasonable, a range was cut out for which linearity seemed to hold good (this usually involved neglecting some of the points for the small values of x). The line of best fit was obtained by the method of least squares and the values in Table 2 calculated. $n \pm$ s.d. (standard deviation) is the slope and a (μ) is called by Best the scale diameter. In 12 cases out of 28, blanks are left where curvature was too great to enforce a straight line on the data. Fig. 5 shows three of the most and three of the least linear plots. Where linearity does hold it is often for a much wider range (more than 1-100 μ) than any considered by Best.

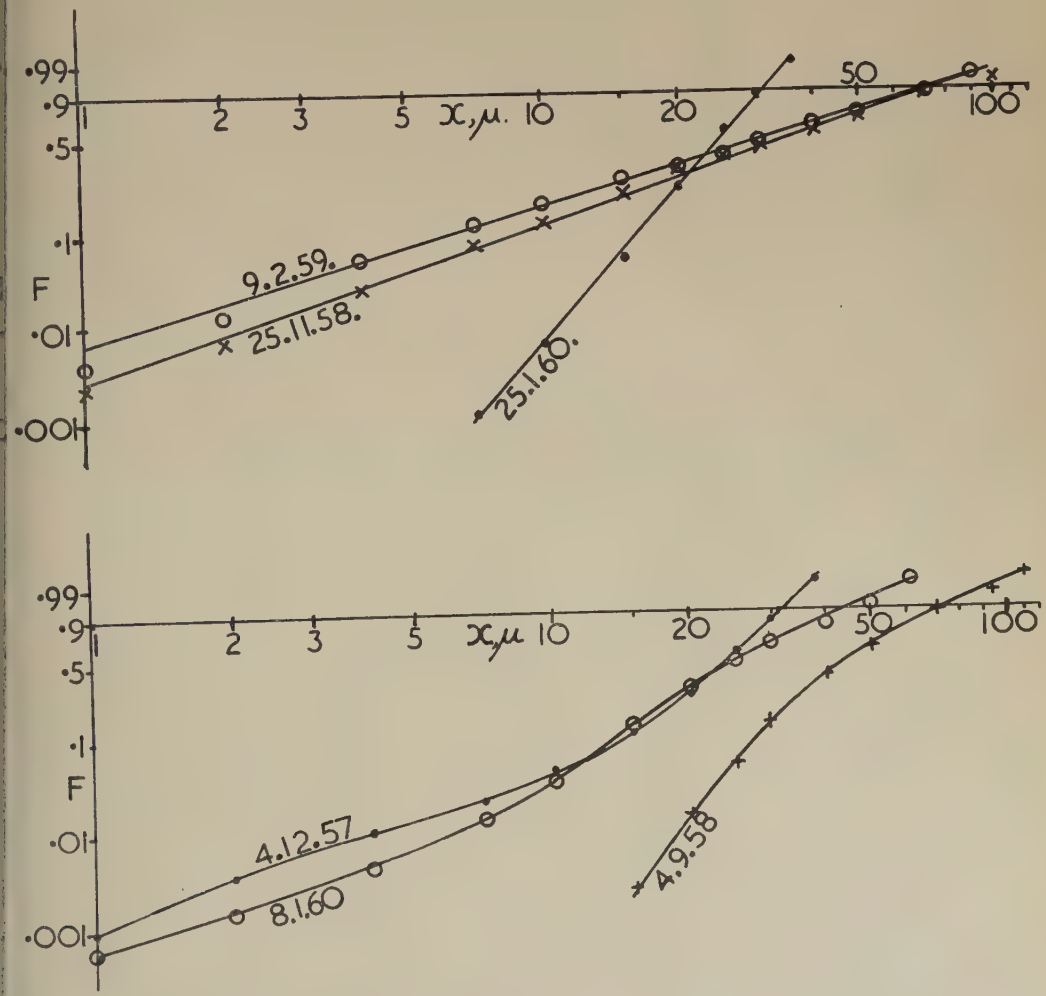

Figure 5. Fraction F of total fog-water volume in drops smaller than x. Log $\log_{10}(1/1 - F)$ plotted against $\log_{10} x$ (Best's equation).

TABLE 2.

(1) Date	(2) Air temp. (°F)	(3) Visual range (m)	(4) Drop diameter range (μ)	(5) Number median diam. (μ)	(6) Volume median diam. (μ)	(7) Mean area diam. (d_a, μ)
1957						
19 February	35	200	1-160	2·2	50	5·6
5 March	41	500	1·6-160	6·6	48	12·8
3 April	—	300	1·2-320	2	85	6
4 December (0900 hr)	29	150	0·8-40	1	24·5	4·3
4 December (1430 hr)	33	120	0·85-55	1·4	31	8·2
1958						
31 January	31	120	0·8-190	1·1	31	5·2
3 March	36	250	1·3-46	9	27	14·5
4 September	57	600	16-120	30	42·5	35
25 November	42	400	0·8-143	0·9	35	3·6
4 December	36	250	0·8-124	1	26	5·7
17 December	42	250	0·8-115	1·05	25	5·5
22 December	42	200	0·8-66	< 1	27	5·4
1959						
29 January	30	80	0·9-40	1	19	4·2
6 February (Noon)	32	250	0·9–15	1·5	7	2·3
6 February (1600 hr)	33	200	0·9-29	1·6	7·2	2·6
9 February	36	250	0·85-108	1	33	3
1960						
4 January	50	400	0·65-65	< 1	26	4·2
8 January	31	350	0·65-72	< 1	27	6·9
25 January	33	160	2·6-42	19·5	24	21
8 September	52	300	0·5-88	0·5	32	3·9
17 October (1000 hr)	—	140	1-66	1·6	40	10·3
17 October (1030 hr)	41	160	1-64	2·7	36	14
31 October	42	170	1·2-54	2·3	24	8·9
9 December (0930 hr)	32	180	0·9-60	1·7	25	9·1
9 December (Noon)	33	260	0·8-77	1·7	29	8
13 December (0930 hr)	30	100	1-36	2·3	19	7·3
13 December (1030 hr)	32	80	1·3-45	2·4	19	4·9
13 December (1230 hr)	33	75	1·5-43	2·3	12·5	4·5

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(10)	(11)			(12)	(13)	(14)	(15)	(16)
Water content (gm m^{-3})				Calculated σ_D from (7) and (9)	Observed σ from (3)	Best's equation		
From impactor	By weighing			Km^{-1}	Km^{-1}	n	$\pm$ s.d.	a
0·130	0·128†,	0·132†		8·2	20	1·54	0·029	79·8
0·0043	0·0044†,	0·0045†		0·37	8	1·64	0·050	63·3
0·046	0·06*			3·1	13	1·53	0·053	112·7
0·058	0·044*,	0·185†		8·8	28			
0·18	0·096*	0·25†		18·8	33			
0·150	0·21*	0·156†		14	33			
0·029	0·078*	0·041†		3·8	16	3·63	0·088	28·9
0·0104		—		0·73	6·5			
0·026		—		3·5	10	1·49	0·030	46·6
0·069	0·10*	0·30†		9·5	16			
0·08		—		12	16			
0·095	0·128*,	0·075†		12	20			
0·143	0·146′			30	50			
0·0015	0·04′			0·9	16			
0·0071	0·05′			3·5	20	1·38	0·058	10
0·040	0·081′			7	16	1·43	0·013	41
0·032		—		4·1	10			
0·092		—		10·8	11·4	2·75	0·050	37
0·13	0·195′			15·6	24	5·58	0·133	26·5
0·072		—		7·7	13	2·54	0·029	37·7
0·297	0·27′			26	28	3·45	0·133	44
0·117	0·17′			12	24	3·40	0·036	39·7
0·17	0·13″			24·5	23	3·05	0·044	27·8
0·145	0·19′	0·16*	0·14″	20	22		—	
0·091	0·095′	0·041*	0·108″	11	15		—	
0·068	0·104′	0·192′	0·13*	13	39	2·82	0·115	22·8
0·158	0·134′	0·19*	0·18*	34	49	2·27	0·068	20·5
0·102		—		31	52	2·18	0·087	16·4

(*b*) *Water-content measurements*

The water-content figures in Cols. 10 and 11, Table 2, show that, in general, there is fair agreement between the two very different methods of weighing and microscope sizing. The microscope method has large potential errors in estimating liquid volume because the measured diameters of the stains have to be converted to drop diameters by a variable calibration factor, and then cubed. It is gratifying, therefore, that there is frequent good agreement. However, on the average, the impactor estimates tend to be lower than the weighted samples, suggesting there may still be a small systematic error in the impactor assessment. This could be an internal wall loss in the intake orifice and elsewhere, which cannot apply to the weighing methods, where all sampled drops are weighed.

Evaluation of the relative merits of the weighing methods has been hampered by the rarity and non-persistence of most of the local morning fogs in which the work had to be done. Usually there was time for only one weighing sample to be taken before the fog cleared or changed. Where there are comparable figures in Col. 11 it is seen that the samples are variable, but whether this is a real effect or is an experimental error is uncertain. The glass impactor (Fig. 3) is certainly the simplest suction device and its samples appear to be as good as any. The grid of wires (Fig. 4) was unsuitable for use in thin or 'dry' (small droplet) fogs such as that of 13 December 1960.

Most of the liquid water-content figures are appreciably lower for similar visibility than those reported by Radford (1938). Size distributions given for Radford's sea fogs suggest they were coarser than the present ones, which could account for the difference. This is speculation, as the drop-size methods cannot be compared.

The worst agreement between the impactor and weighing methods is in the two samples of 6 February 1959 which were taken in a rare persistent fog which proved to be smog. There was very little liquid water in the samples, and much of that was little more than a film around dense black nuclei. This no doubt is typical of the appearance of many town fogs and the airmass appears to have drifted down from the industrial midlands on this occasion. The filter samples in Col. 11 would give some measure of the total mass of airborne pollution, which would give most of the opacity in this case, while the figures in Col. 10 show the low liquid-water content. On a few occasions black, smoke-like, nuclei have been observed in the drop impressions on the gelatin slides. When this occurs there is a correlation (with much scatter) between the number of discrete particles seen per drop and the drop size, strongly suggesting growth by frequent collisions even down to about 5 μ, below which there is usually only a single nucleus. If this is so, it is in direct conflict with the calculations of Hocking (1959) which gave no collisions for drops below 36 μ diam. In general, droplets from the local country fogs are clean in the sense that they leave no solid residue visible under the high-power microscope.

(*c*) *Attenuation*

Reference to Cols. 12 and 13 of Table 2 shows that agreement between the scattering coefficients calculated from the impactor counts and the observed attenuation coefficients from Koschmieder's formula, vary from very good to very poor. The calculated figures are particularly low for the thin fogs of 5 March 1957 and 4 September 1958, also for the smog of 6 February 1959, but the high solid-particle content of the latter is not inconsistent with a low water-scattering value. After 25 November 1958, when the technique had been improved to its final form, the average value of Col. 12 and Col. 13 is 0·65 (neglecting 6 February 1959). This is a substantial agreement between visibility theory and impactor sample calculations, particularly when compared with similar work (e.g., Elton *et al.* 1956); (Stewart 1957), but it is highly significant that the impactor samples are lower in every case except one. Possible causes for this are :

(*i*) *Errors in the optical theory.* The droplet scattering theory used for Col. 12 is too firmly established to be questioned (Middleton 1952) but this is not so for Koschmieder's

quation used for Col. 13. The basis here is ϵ, the brightness difference threshold for e eye, which is taken as 0·02. As shown by Middleton (1952 p. 219) higher values of is have been found, but to reduce the Col. 13 figures to the average of Col. 12, ϵ would ave to be 0·075 which is certainly too high. In the present experiments if ϵ were as high 0·03, which is possible, the Col. 13 figures would only be reduced by 10 per cent.

(*ii*) *Attenuation due to particles other than water drops.* For nearly all the present untry fogs this is certainly negligible, because visibility goes up at once to several miles hen the fogs follow their usual course of mid-morning evaporation. This would not be for town fogs where there can be much smoke, so that differences between Cols. 12 nd 13 would then be expected.

(*iii*) *Underestimation of drop size.* To raise Col. 12 to average parity with Col. 13, stimated drop diameters would have to be increased by 25 per cent. In the ranges of rop size containing most of the volume and area of fog water, diameters of stains on he gelatin film can be measured to ± 5 per cent or better and the calibration against agnesium oxide should be accurate to better than 10 per cent. Therefore the factor ought is apparently beyond the precision of the measurements. Also, a 25 per cent ncrease in diameter would double the Col. 10 figures of water content which would then ertainly be too high.

(*iv*) *Sampling errors.* It is likely that some of the difference between Cols. 12 nd 13 is still due to the incomplete elimination of the errors described in Section 3. t is particularly noteworthy that if the drop-size distributions obtained from the impactor re applied to the weighed water-content figures of Col. 11, instead of Col. 10, then greement between Cols. 12 and 13 would, on the average, be close. This suggests that ny loss of drops which may occur inside the impactor does not have a serious effect on he size distribution. Deficiencies in Col. 12 could then be attributed to the Col. 9 figures eing lower than reality.

Drops of about 1 μ diameter merit special consideration, for it is near this size that he scattering-area ratio attains a sharp peak value of nearly 4, instead of the value 2 sed in all the present calculations. The difference between Cols. 12 and 13 might be ntirely due to a serious error in sampling these small drops. However, although drops f about 1 μ were often present in very large numbers their contribution to the total ross-section area of the whole spectrum of drops was always very small. The maximum alue found was 9 per cent of the total on 9 February 1959; usually it was very much lower han this. In fogs where Cols. 12 and 13 are markedly different the average value of the ercentage of total area contributed by drops in the range 0·8 μ-1·5 μ is 2·6 and at a alue for K of 3·5 for these drops, an average 20-fold increase in the number sampled would be required to raise Col. 12 to parity with Col. 13. It is true that at 1 μ no great ccuracy of sampling can be claimed because the drop retention efficiency of the apparatus described is only 15 per cent at 1 μ so that the numbers counted had to be increased 7-fold to give the estimated numbers. Efforts were made to sample drops of 1 μ and smaller with high accuracy by using an isolated stage 4 Casella impactor, with reciprocating slide, which had nearly 100 per cent impaction efficiency at 1 μ and allows no chance of droplet evaporation before impact. Invariably this device failed to reveal more 1 μ drops than the standard method. The gelatin film is fully sensitive at this size and on one occasion only (8 September 1960), drops down to 0·5 μ were found by the isolated stage 4. The evidence from the present work therefore, is that the large numbers of very small drops in the local fogs make only a minor contribution to the opacity.

(*v*) *Errors in visual range estimation by eye.* This is notoriously uncertain and reasonable accuracy can only be claimed in the present work for distances up to 250 m. Beyond that, landmarks such as clumps of trees or buildings were observed but the writer does not feel that the large underestimates of visual range which would be required to reduce the higher values of Col. 13/Col. 12 are possible.

To conclude, discrepancies between calculated and observed values of attenuation coefficients, where they occur, may largely be explained by the causes given, but the writer is not satisfied that this is adequate to explain the large discrepancies which were consistently found for thin fogs, and which remain, at present, unaccountable. The rarity, transience and rapid variability of most of the local fogs, together with many sources of experimental error, have made an overall estimate of the precision of the results impossible. Nevertheless, the results are probably the most complete which have been presented for fogs, and the frequent good agreement between observed and estimated opacity has never previously been obtained. The method should yield profitable results in clouds.

8. Acknowledgments

I gratefully acknowledge the assistance of Mr. S. Peto and his staff with the statistical computations, and of Mr. R. Clifford with the field work. Valuable criticism was given by Professor B. J. Mason.

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Statistics of lateral gustiness at 16 m above ground

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(Manuscript received 21 December 1960; in revised form 12 June 1961)

SUMMARY

The horizontal lateral wind-fluctuations at 16 metres above ground have been recorded over periods totalling 1,350 hours by a system described by Jones and Pasquill (1959). Hourly-average values Θ_τ of the standard deviation of wind fluctuations sampled within times $\tau = 5$, 30, and 180 sec are classified according to wind speed and stability $s = \Delta T/U^2$ °F/(m/sec)^2 (where ΔT is the temperature difference between 7·1 m and 1·2 m, U is the wind speed at 15·5 m). A critical value of stability $s = +0{\cdot}017$ is indicated separating near-neutral or unstable stabilities ($s < 0{\cdot}017$), for which Θ_τ depends primarily on stability, from very stable conditions ($s > 0{\cdot}017$), for which wind speed is the dominant parameter. Some support for this transition is also found in the variation of spectrum shape with stability. A second transition on the unstable side may correspond to Priestley's critical value separating forced- and free-convection regimes.

The results are in a form suitable for application to particle diffusion in the atmosphere. However, the large variations in wind direction in very stable conditions with light winds cannot be interpreted, at present, in terms of diffusion with the same degree of certainty as in near-neutral stabilities.

1. INTRODUCTION

Data on the variation of horizontal lateral atmospheric turbulence with wind speed and stability and their relation to lateral diffusion near the ground, have been obtained. A lightweight high-response wind-vane, operating a sensitive micro-torque potentiometer (net resolution time about 0·2 sec), was mounted at 16 m on a tower situated on gently rolling downland to measure the wind fluctuations. Over 1,000 hours of such records, taken during two 3-monthly periods in the winter and summer of 1959, have been analysed.

2. DATA

Using the equipment described by Jones and Pasquill (1959), the wind fluctuations were presented on three slowly-moving charts ($\frac{1}{4}$ in. hr^{-1}) as smoothed values of the r.m.s. (standard deviation) angular fluctuation Θ_τ of the vane when the fluctuations were sampled over durations τ equal to 5, 30 and 180 sec. The mean values of Θ_τ appropriate to consecutive hourly intervals, were read off the trace to within an estimated accuracy of 2 per cent, approximately constant since the variability increases with increasing Θ_τ. Corresponding wind speeds were determined from a rotating-cup anemometer mounted slightly below the vane at 15·5 m on the same tower. A temperature difference between 1·2 m and 7·1 m, was measured to within about 0·1°F (except for very erratic traces) by permanent apparatus fixed to the tower.

Finally, the wind direction was estimated to the nearest 10° from a standard Baxendell recorder at a height of 13 m situated about 100 m away. Completely unobstructed fetch exists only when the wind direction lies between 040°-180°-260°, and the basic data have been restricted to this sector.

The periods analysed are given in Table 1.

Although 1,350 hours were taken from these periods, complete generality cannot be expected; for example, more stable periods occurred during winter than summer. The selective analysis of the summer periods was an attempt to equalize the numbers of hours with stable, neutral and lapse conditions to give a fairly balanced description of the relation between turbulence and stability.

TABLE 1. THE PERIODS ANALYSED

Period (1959)	Number of days	Remarks
13-18 January	6	
31 January-7 February	8	
11 February-1 March	19	All analysed
7 March-12 March	6	
21 March-31 March	11	
19-21 June	3	
23 June-7 July	15	Day-time + 12
10-24 July	15	nights analysed
26 July-16 August	22	

Distributions of the number of times a given value of Θ_τ occurred for specified τ are quite markedly peaked and show no appreciable difference between summer and winter data, implying that selection introduced no great bias.

Table 2 shows that the percentage ratio of standard deviation to mean (the coefficient of variation) increases as τ increases; in other words, the low-frequency components of the turbulence are influenced proportionally to a greater extent than the high-frequency components by variations in stability and wind speed. This variation will become more apparent in later sections.

TABLE 2. THE MAIN PARAMETERS OF THE FREQUENCY-DISTRIBUTION OF Θ_τ IN DEGREES

τ (sec)	Mode	Mean	Standard deviation
5	3·15	3·08	0·94
30	4·65	4·70	1·80
180	5·85	6·96	3·70

3. STABILITY

In theoretical investigations the usual criterion of stability is the Richardson number

$$Ri = \frac{g}{T}\frac{d\theta}{dz}\left(\frac{dU}{dz}\right)^{-2}$$

where θ = potential temperature, T = absolute temperature and U = mean wind speed. In practice this is often simplified, by assuming that the gradient of wind is proportional to the speed, to

$$s = \frac{T(z_1) - T(z_2)}{U^2(z_3)}\ ^\circ\mathrm{F}/(\mathrm{m/sec})^2,\ z_1 > z_2$$

In the present work, $z_1 = 7{\cdot}1$ m, $z_2 = 1{\cdot}2$ m and $z_3 = 15{\cdot}5$ m. Full interpretation and comparison of results with those obtained elsewhere are complicated by differences in z_1, z_2 and z_3, and by the complex dependence of turbulence on the whole stability distribution with height.

Attempting to relate the present form of s with the more general Richardson number, previous data on the detailed wind and temperature profiles at lower levels on the same site were analysed, and by assuming that Ri varies linearly with height, an approximate relationship was deduced for its value at 16 m,

$$|Ri| = (10\,|s|)^{3/2}$$

This approximate relation was deduced from negative and near-zero values of s only, but may be used in rough comparisons with the results of other trials when the relevant range of Ri varies over several orders of magnitude.

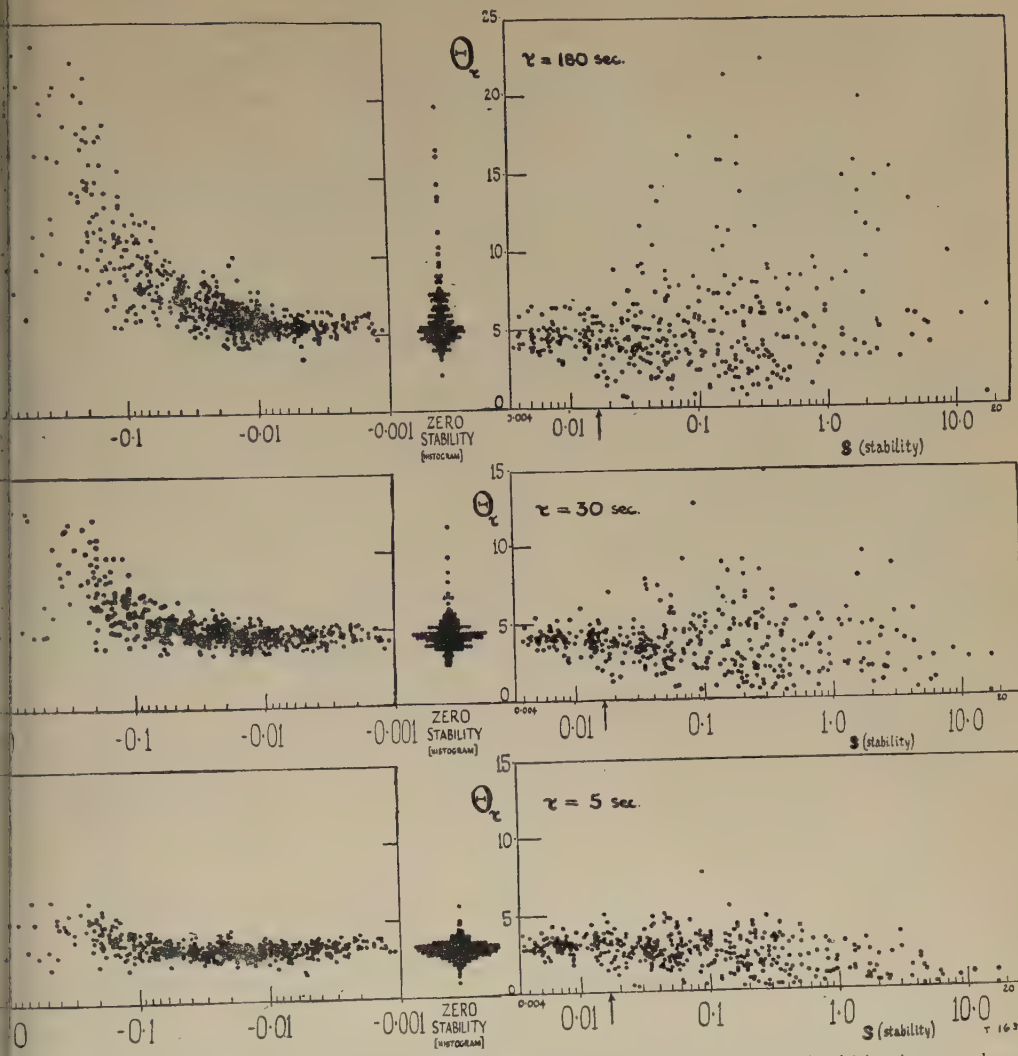


Figure 1. The standard deviation of horizontal lateral wind fluctuations, sampled within time τ, plotted against stability, s.

4. Effect of stability on the value of Θ_τ

On investigation it appeared that for large τ, the average value of Θ tended to increase at low wind speeds, owing largely to stable night-time values. This stimulated an analysis of the data principally in terms of stability. Fig. 1 shows all the values of hourly-mean Θ_τ plotted against log $|s|$ for three values of $\tau = 5$, 30 and 180 sec and for s positive and s negative. The values for zero stability are shown as histograms where the abscissae represent the number of observations.

Rather more scatter is immediately noticeable in these values, owing almost entirely to cases of low wind speed. As such they represent the least reliable section of all the data for two reasons : firstly, the wind speed changes during the hourly averaging period and may become zero at times. This means that the stability, inversely proportional to the square of the wind, varies considerably and has an average value much greater than s, obtained from *mean* values of wind speed and temperature gradient. Secondly, since average values of ΔT are known only to within 0·1°F, s is comparatively uncertain at low wind speeds. It therefore seems possible that these abnormal points at $s = 0$ are more typical of values of Θ_τ at large positive and negative values of s.

Some points of interest are indicated by these distributions :

(*i*) Generally, stability alone is not a very satisfactory measure of Θ_τ.

(*ii*) There is rather less scatter for lapse conditions ($s < 0$) than for inversio conditions ($s > 0$).

(*iii*) For $s < 0$ the transition from small scatter to larger scatter coincides with a upward trend in Θ_τ.

(*iv*) This transition is not very sharp but apparently occurs at different values of for different values of τ; (see Table 3).

(*v*) It is interesting to compare the values of Ri in Table 3 with the narrow rang $-\cdot02 \leqslant Ri \leqslant -\cdot05$ suggested by Priestley (1955) as marking the transition for *vertic* turbulence from dominantly forced convection to free convection. The extent to whic these two phenomena are related is at present uncertain.

(*vi*) There is a reasonable continuity in the general trend of points through zer stability.

(*vii*) On the stable side, the scatter remains small up to a region of s, 0·01 to 0·0 (see Table 4). Beyond, the scatter is very much increased, especially for Θ_{180}, and stabilit is a very poor indicator of the values of Θ. A second transitional, or *critical*, value of is therefore indicated almost independent of τ, best estimated at about 0·017.

TABLE 3. TRANSITIONAL VALUES FOR NEGATIVE STABILITY

τ (sec)	Transitional range of s	Corresponding Ri
5	− 0·06 to − 0·10	− 0·5 to − 1·0
30	− 0·02 to − 0·06	− 0·1 to − 0·5
180	− 0·006 to − 0·01	− 0·02 to − 0·04

TABLE 4. STANDARD DEVIATION (SCATTER) OF THE INDIVIDUAL HOURLY AVERAGES OF Θ_τ

S	From		−0·16	−0·08	−0·04	−0·02	−0·01	0·004	0·01	0·017	0·04	0·08	0·16	
	To	<−0·16	−0·08	−0·04	−0·02	−0·01	+0·01	0·01	0·017	0·04	0·08	0·16	0·32	>0·32
τ	180	4·47	2·67	1·65	1·14	0·94	1·73	0·69	0·80	1·77	3·35	3·18	4·44	3·80
(sec)	30	2·37	1·25	0·71	0·61	0·56	0·89	0·24	0·63	1·22	2·11	1·81	2·78	2·01
	5	1·11	0·68	0·38	0·46	0·41	0·52	0·41	0·53	0·78	1·10	1·00	1·54	1·08

5. THE CRITICAL STABILITY

The existence of a critical positive stability may imply that vertical turbulence is either damped out almost completely or is no longer the main source of energy for the horizontal fluctuations concerned in Θ_{180}. These may come from spatial variations in density and temperature (i.e., of thermal origin) rather than from dynamic origin (i.e., resulting from vertical wind shear). Beyond the critical stability an alternative parameter by which Θ may be more closely defined appears to be the velocity. The high values

Θ are generally associated with low wind speeds and values of Θ for high wind speeds e generally low. However, the parameter is insufficient in so far as the scatter of individual lues remains high, particularly in the low wind-speed bands. All the data have been ouped according to wind speed and stability and the values of Θ_5, Θ_{30} and Θ_{180} averaged each group (see Table 5). Because of imperfect vane and recorder response Θ lacks comparatively small contribution from the high-frequency eddies (the effect is roughly quivalent to averaging the full fluctuations over running periods of 0·2 sec). So as to btain a value of Θ with unambiguous meaning, values of $\sqrt{(\Theta_{180}^2 - \Theta_5^2)}$ were formed om the group-averages and plotted against stability with a different symbol for each band wind speed. Stability grouping is the same as in Table 5, except that the $-0{\cdot}01$ to $0{\cdot}01$ is subdivided into three bands ($-0{\cdot}01$ to $-0{\cdot}001$; 0; and $+0{\cdot}004$ to $0{\cdot}01$).

The results are shown in Fig. 2. The bracketed symbols contain less than five hours f data and therefore lack the significance of others, many of which contain twenty or ore hours. Again the values for $s = 0$, and low wind speeds must be treated with caution s explained in the previous section.

Mean lines for each wind-speed band have been drawn by eye on Fig. 2. The existence f a transition at or near $s = 0{\cdot}017$ is shown by the marked separation of the wind-speed nes on the stable side and their apparent collapse to a single line on the other.

Since this interpretation requires the high values at $s = 0$ to be ignored, further vidence for the existence of a critical stability must be examined. From Table 5, curves f Θ_τ against u may be drawn for constant values of s. For s negative and for $s < 0{\cdot}017$, Θ characteristically increases slightly or remains constant as u increases (except for small alues of u). A pilot investigation has indicated that if the period of averaging is reduced rom 1 hr to 15 min, to counteract the distortion discussed in the previous section, this haracteristic applies to all values of U. In contrast the curves for $s > 0{\cdot}017$ all show distinctive decrease of Θ with increasing u.

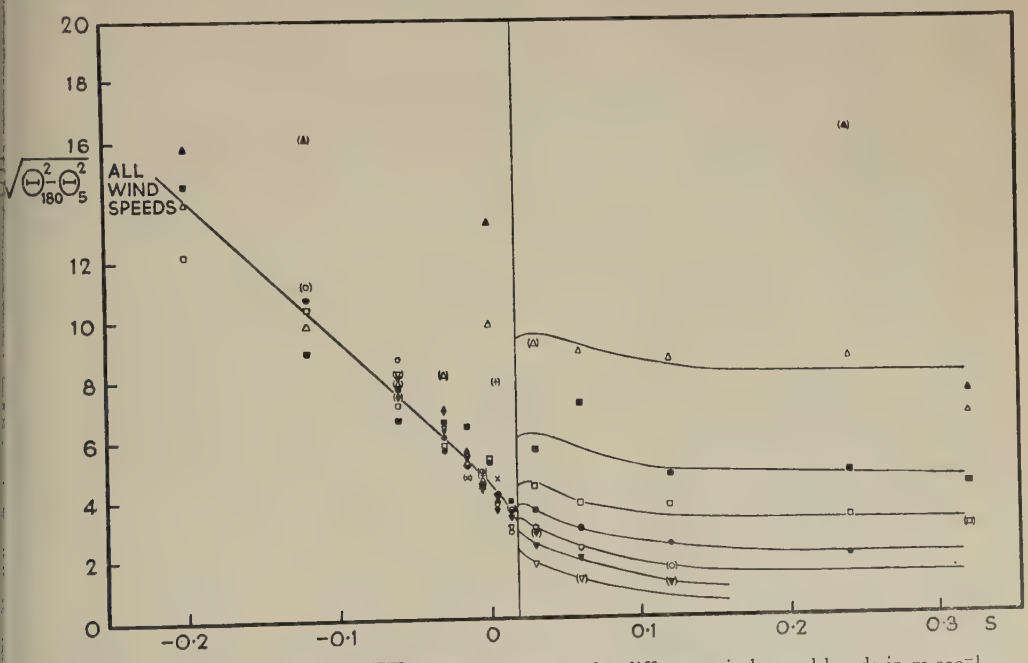


Figure 2. Values of $\sqrt{\Theta_{180}^2 - \Theta_5^2}$ against stability s, for different wind-speed bands in m sec^{-1}.

▲ = 0-1 □ = 3-4 ▼ = 6-7 ◇ = 9-10 ◑ = 12-13
△ = 1-2 ● = 4-5 ▽ = 7-8 × = 10-11
■ = 2-3 ○ = 5-6 ◆ = 8-9 + = 11-12

(Symbols bracketed where there are less than 5 cases)

TABLE 5. MEAN VALUES OF Θ_5, Θ_{30}, Θ_{180} IN DEGREES AND p FOR VARIOUS STABILITY AND WIND-SPEED BANDS

σ = Standard deviation of individual hourly averages contributing to Θ, suffixes refer to sampling time in seconds
N = Number of hourly averages contributing to Θ
$\bar{u}$ = Mean of hourly averages of wind speed in m sec^{-1}
p is defined by $G(n) \propto n^{-p}$ (see text)

Stability (s) range	$< -0{\cdot}16$						$-0{\cdot}16$ to $-0{\cdot}08$						$-0{\cdot}08$ to $-0{\cdot}04$					
Wind-speed range	N	$\bar{u}$	Θ_5 (σ_5)	Θ_{30} (σ_{30})	Θ_{180} (σ_{180})	p	N	$\bar{u}$	Θ_5 (σ_5)	Θ_{30} (σ_{30})	Θ_{180} (σ_{180})	p	N	$\bar{u}$	Θ_5 (σ_5)	Θ_{30} (σ_{30})	Θ_{180} (σ_{180})	p
0·0-0·9	23	0·68	3·98 (1·37)	8·42 (3·73)	16·28 (4·47)	1·70	1	0·8	3·8	10·2	> 16·5	1·35						
1·0-1·9	36	1·43	4·51 (1·07)	8·43 (2·29)	14·61 (4·79)	1·58	16	1·58	3·14 (0·55)	5·87 (0·91)	10·28 (1·86)	1·59	4	1·63	2·85	5·13	8·37	1·49
2·0-2·9	18	2·51	4·87 (0·92)	8·51 (1·93)	15·30 (4·24)	1·67	29	2·48	3·72 (0·67)	5·98 (1·35)	9·62 (2·67)	1·53	10	2·49	3·00 (0·39)	4·89 (0·75)	7·23 (1·33)	1·65
3·0-3·9	10	3·21	4·44 (0·58)	7·68 (1·41)	12·90 (3·04)	1·56	28	3·44	4·01 (0·70)	6·60 (1·38)	11·09 (3·22)	1·59	21	3·50	3·22 (0·34)	5·10 (0·67)	7·78 (1·46)	1·44
4·0-4·9							17	4·42	4·11 (0·28)	6·59 (0·50)	11·42 (1·59)	1·66	25	4·54	3·33 (0·34)	5·24 (0·67)	8·34 (1·39)	1·54
5·0-5·9							3	5·20	3·97	6·40	11·80	1·76	23	5·40	3·54 (0·37)	5·52 (0·73)	9·32 (1·77)	1·64
6·0-6·9													9	6·24	3·44 (0·17)	5·39 (0·31)	8·68 (1·19)	1·55
7·0-7·9													2	7·80	3·35	5·35	8·80	1·58
8·0-8·9													1	8·0	3·3	5·1	8·1	1·54

Stability (s) range			− 0·04 to − 0·02						− 0·02 to − 0·01						− 0·01 to + 0·01				
Wind-speed range	N	$\bar{u}$	Θ_5 (σ_5)	Θ_{30} (σ_{30})	Θ_{180} (σ_{180})	p	N	$\bar{u}$	Θ_5 (σ_5)	Θ_{30} (σ_{30})	Θ_{180} (σ_{180})	p	N	$\bar{u}$	Θ_5 (σ_5)	Θ_{30} (σ_{30})	Θ_{180} (σ_{180})	p
0·0-0·9													6	0·63	2·28 (0·96)	5·93 (2·64)	13·33 (4·57)	1·87
1·0-1·9	3	1·50	2·63	4·60	8·47	1·71							6	1·57	3·60 (1·36)	6·70 (2·63)	10·40 (4·50)	1·38
2·0-2·9	7	2·49	2·93 (0·87)	4·73 (1·10)	7·16 (0·90)	1·41	7	2·69	2·80 (0·60)	4·61 (1·03	6·93 (1·53)	1·39	16	2·55	2·83 (0·84)	4·37 (1·43)	5·87 (1·75)	1·18
3·0-3·9	20	3·45	2·87 (0·36)	4·37 (0·60)	6·41 (1·15)	1·39	27	3·53	2·95 (0·47)	4·40 (0·70)	5·89 (1·24)	1·20	59	3·46	2·95 (0·61)	4·45 (0·95)	6·02 (1·62)	1·22
4·0-4·9	38	4·45	3·05 (0·48)	4·46 (0·79)	6·34 (1·00)	1·36	22	4·45	3·10 (0·43)	4·49 (0·62)	5·90 (0·88)	1·18	51	4·47	2·92 (0·36)	4·11 (0·59)	5·10 (0·93)	1·05
5·0-5·9	29	5·43	3·13 (0·40)	4·71 (0·15)	6·78 (1·04)	1·36	17	5·39	3·10 (0·38)	4·48 (0·50)	5·93 (0·68)	1·20	37	5·46	3·16 (0·38)	4·32 (0·65)	5·15 (0·99)	0·95
6·0-6·9	28	6·36	3·25 (0·37)	4·78 (0·52)	7·01 (1·04)	1·42	20	6·34	3·33 (0·38)	4·75 (0·58)	6·27 (0·89)	1·21	43	6·39	3·09 (0·45)	4·25 (0·57)	5·05 (0·77)	0·92
7·0-7·9	15	7·30	3·34 (0·28)	4·87 (0·38	7·13 (0·73)	1·43	22	7·52	3·32 (0·22)	4·72 (0·33)	6·34 (0·59)	1·26	47	7·42	3·37 (0·44)	4·59 (0·59)	5·46 (0·67)	0·94
8·0-8·9	18	8·36	3·52 (0·34)	5·08 (0·43)	7·78 (0·70)	1·53	28	8·38	3·41 (0·33)	4·76 (0·42)	6·47 (0·75)	1·31	51	8·44	3·35 (0·29)	4·51 (0·43)	5·50 (0·76)	0·95
9·0-9·9							10	9·24	3·18 (0·34)	4·50 (0·41)	6·26 (0·88)	1·35	41	9·41	3·19 (0·35)	4·33 (0·42)	5·23 (0·61)	1·00
10·0-10·9							3	10·30	3·30	4·50	5·70	1·15	19	10·46	3·31 (0·21)	4·40 (0·33)	5·53 (0·85)	1·16
11·0-11·9													4	11·33	3·23	4·80	6·43	1·21
12·0-12·9													1	12·2	3·2	4·3	5·8	1·34

TABLE 5 (*continuation*)

σ = Standard deviation of individual hourly averages contributing to Θ, suffixes refer to sampling time in seconds
N = Number of hourly averages contributing to Θ
$\bar{u}$ = Mean of hourly averages of wind speed in m sec^{-1}
p is defined by $G(n) \propto n^{-p}$ (see text)

Stability (s) range	+ 0·010 to + 0·017						+ 0·017 to + 0·04						+ 0·04 to + 0·08					
Wind-speed range	N	$\bar{u}$	Θ_5 (σ_5)	Θ_{30} (σ_{30})	Θ_{180} (σ_{180})	p	N	$\bar{u}$	Θ_5 (σ_5)	Θ_{30} (σ_{30})	Θ_{180} (σ_{180})	p	N	$\bar{u}$	Θ_5 (σ_5)	Θ_{30} (σ_{30})	Θ_{180} (σ_{180})	p
1·0-1·9							2	1·80	3·30	6·10	9·65	1·42	9	1·57	3·21 (1·47)	5·98 (3·48)	9·39 (5·17)	1·40
2·0-2·9	5	2·78	2·50 (0·51)	3·66 (0·65)	4·60 (1·13)	1·05	6	2·53	3·03 (0·90)	4·80 (1·65)	6·32 (1·86)	1·11	8	2·55	3·34 (1·35)	5·44 (2·57)	7·85 (4·31)	1·31
3·0-3·9	3	3·67	2·77	3·77	4·53	0·98	16	3·44	2·80 (0·71)	4·11 (1·13)	5·14 (1·83)	1·03	20	3·49	2·97 (0·87)	3·97 (1·28)	4·81 (1·57)	1·03
4·0-4·9	8	4·41	2·83 (0·26)	3·73 (0·32)	4·51 (0·56)	1·05	13	4·52	2·75 (0·46)	3·67 (0·66)	4·46 (1·04)	1·05	16	4·40	2·53 (0·87)	3·25 (1·09)	3·84 (1·10)	1·03
5·0-5·9	10	5·66	2·73 (0·72)	3·56 (0·85)	3·91 (0·90)	0·61	21	5·38	2·72 (0·70)	3·48 (0·66)	4·02 (0·79)	0·92	9	5·37	2·02 (0·82)	2·46 (0·87)	3·01 (0·92)	1·24
6·0-6·9	12	6·36	3·17 (0·38)	4·04 (0·63)	4·57 (0·74)	0·82	16	6·36	2.39 (0·91)	3·01 (1·10)	3·36 (1·19)	0·77	5	6·16	1·74 (0·69)	2·24 (1·00)	2·60 (1·17)	0·92
7·0-7·9	5	7·42	3·42 (0·28)	4·12 (0·34)	4·52 (0·49)	0·76	7	7·34	1·97 (0·97)	2·34 (1·23)	2·63 (1·28)	0·95	3	7·03	1·50	1·80	1·93	0·45
8·0-8·9							3	8·50	3·07	3·87	4·20	0·59						

Stability (s) range	+ 0·08 to + 0·16						+ 0·16 to + 0·32						> 0·32					
Wind-speed range	N	$\bar{u}$	Θ_5 (σ_5)	Θ_{30} (σ_{30})	Θ_{180} (σ_{180})	p	N	$\bar{u}$	Θ_5 (σ_5)	Θ_{30} (σ_{30})	Θ_{180} (σ_{180})	p	N	$\bar{u}$	Θ_5 (σ_5)	Θ_{30} (σ_{30})	Θ_{180} (σ_{180})	p
0·0-0·9							2	0·70	3·30	8·55	16·60	1·66	11	0·66	0·96 (0·70)	2·56 (1·78)	7·46 (4·18)	2·21
1·0-1·9	9	1·58	3·06 (1·21)	5·66 (2·29)	9·07 (4·72)	1·44	9	1·44	3·67 (2·24)	6·44 (3·87)	9·29 (5·47)	1·26	29	1·41	1·62 (0·98)	3·29 (1·45)	6·81 (4·34)	1·82
2·0-2·9	10	2·43	2·69 (0·87)	4·05 (1·31)	5·39 (1·85)	1·18	15	2·34	2·41 (1·62)	3·63 (2·42)	5·32 (3·20)	1·40	22	2·33	2·11 (1·23)	3·31 (1·93)	4·87 (2·42)	1·38
3·0-3·9	13	3·58	2·72 (0·79)	3·62 (1·13)	4·57 (1·53)	1·17	24	3·28	1·93 (1·26)	2·72 (1·93)	3·74 (3·16)	1·33	4	3·10	3·10	1·15	2·73	1·86
4·0-4·9	14	4·39	1·96 (0·58)	2·56 (0·79)	3·06 (0·78)	1·02	6	4·18	1·55 (0·32)	1·92 (0·30)	2·47 (0·47)	1·35						
5·0-5·9	4	5·45	1·03	1·33	1·85	1·47												
6·0-6·9	1	6·4	0·9	1·2	1·4	0·89												

Further evidence is contained in the implied spectrum-shapes in the frequency bands affecting the values of Θ_5, Θ_{30} and Θ_{180}. If the spectra conform to a simple power law

$$G(n) \propto n^{-p}$$

then, as shown by Smith (1961),

$$p = 1 + \left\{ \log \left[\frac{\Theta_{180}^2 - \Theta_{30}^2}{\Theta_{30}^2 - \Theta_5^2} \right] \Big/ \log 6 \right\}$$

If the band of frequencies is within the so-called ' inertial sub-range ' p would take the value 5/3. But if the scale of turbulence is so small that part of the energy is being fed into the system at frequencies affecting Θ_{180}, p will differ from 5/3. The values of p derived from the present data (see Table 5 and Fig. 3) clearly indicate a transition, but one which occurs at a stability which is not so well-defined as that for $\sqrt{(\Theta_{180}^2 - \Theta_5^2)}$. Fig. 3 is obtained by constructing curves for each u-band and then smoothing these into a consistent set. For all wind speeds in very unstable conditions ($s \ll 0$) p is about 5/3 and then falls, by an amount increasing with increasing wind speed, over a fairly wide band about $s = 0$. This is followed by a gradual increase in the value of p. Here again, on the stable side $s > 0$, wind speed seems a more significant parameter than stability, although stability still appears to matter.

To avoid confusion, the basic points for only two of the wind-speed curves are given in the figure. It may be noted that many of the data in the low wind-speed bands are included at the large $|s|$ parts of the curves, whereas the data in the high wind-speed bands are grouped near $s = 0$. The reliability of the curves thus varies with position.

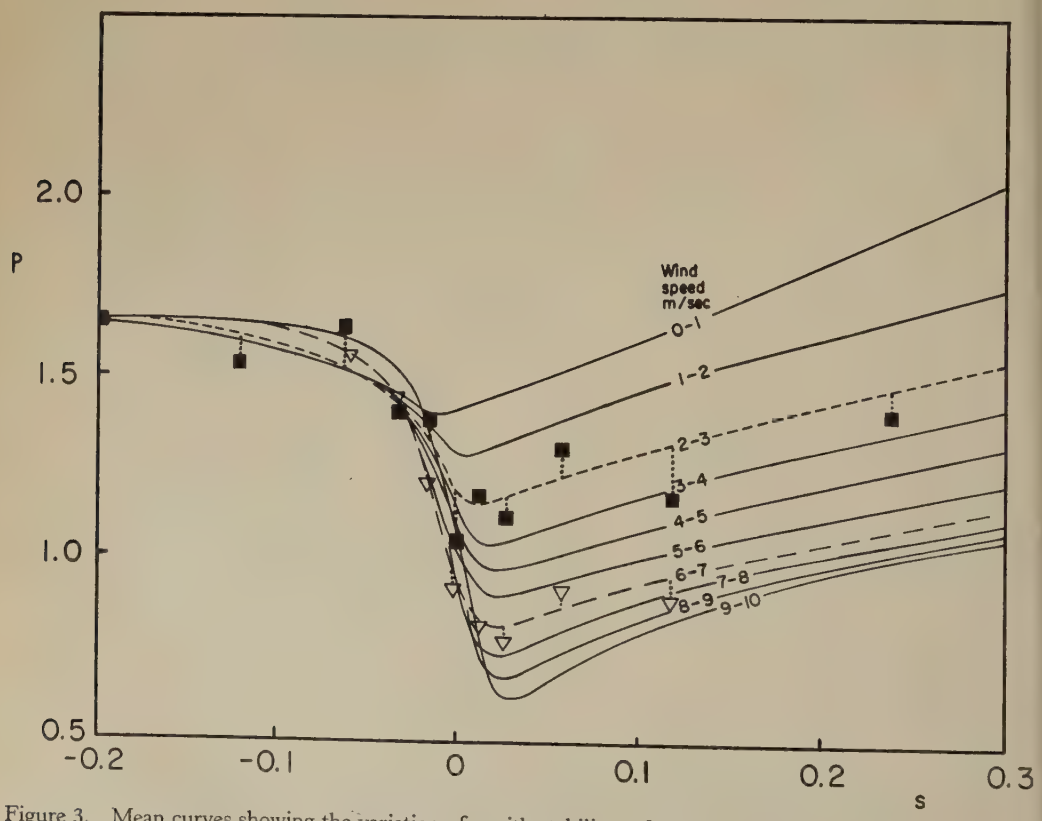


Figure 3. Mean curves showing the variation of p with stability, s for different ranges of wind speed within the 5 sec, 30 sec and 180 sec band.

TABLE 6. STABILITY VALUES FOR MINIMUM p, CLASSIFIED ACCORDING TO WIND SPEED

U (m sec^{-1})	0-1	1-2	2-3	3-4	4-5	5-6	6-7	7-8	8-9	9-10
S	−0·01	+0·005	+0·011	+0·017	+0·021	+0·022	+0·022	+0·023	+0·027	+0·030

Apart from the lowest wind-speed band, which is rather doubtful because of the small number of values involved, the trough-values of s are not greatly different from the previously derived critical value of 0·017 (see Table 6). However, some systematic variation is apparent, and this may indicate a corresponding variation in the critical value largely masked in the earlier analysis.

These values of p imply that the time-scale of turbulence takes on the smallest values near neutral conditions and also for largest wind speeds. On the lapse side, the increase in the effective scale may be due to the addition of convective turbulence, whilst on the stable side the large values may possibly indicate that dynamic turbulence has been replaced by motions originating from horizontal variations of temperature.

6. VARIATION OF Θ WITH τ

Classifying Θ_τ according to wind speed or stability only goes some way to reducing the scatter of values about the mean when the variation of Θ with sampling time τ is considered. For $s < 0{\cdot}017$, this scatter is roughly indicated by the width of the $\Theta - s$ distribution (Fig. 1) at any given value of s. A more direct guide is provided by Fig. 4, which shows

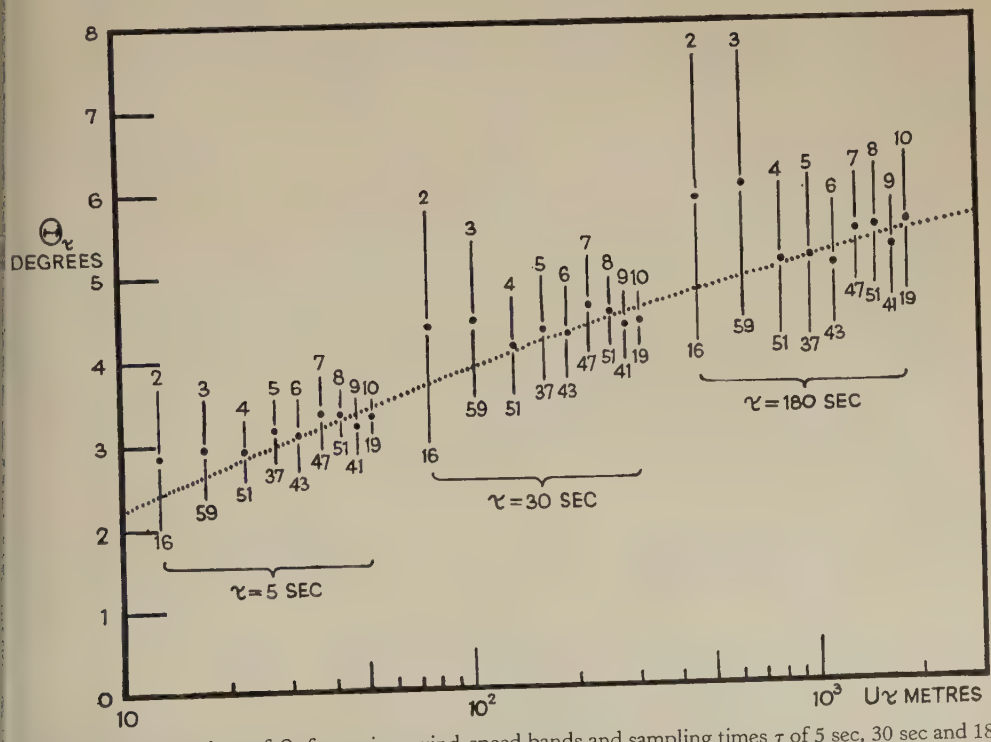


Figure 4. Mean values of Θ_τ for various wind-speed bands and sampling times τ of 5 sec, 30 sec and 180 sec, plotted against run of wind U_τ during neutral conditions ($-0{\cdot}01 \leqslant s \leqslant 0{\cdot}01$). The dotted curve has been drawn through points where $U \geqslant 4$ m sec^{-1}. Vertical lines are the limits of standard deviation, σ from the mean values of Θ_τ denoted by black dots. The figures above the vertical lines indicate the wind-speed bands, i.e., 2 is the range 2·0-2·9, 3 the range 3·0-3·9, etc. The number of individual hourly averages N contributing to Θ_τ is placed below the vertical line.

the mean value of Θ plotted against $U\tau$ for neutral conditions $|s| \leqslant 0{\cdot}01$, for the different wind-speed bands, with vertical lines to denote the scatter ($\pm$ s.d.). The mean curves of Θ against τ are presented in Figs. 5 and 6 for other stabilities. Fig. 5 gives curves for eight stability bands, near-neutral or unstable, below $s = 0{\cdot}017$; the most unstable band, $s < -0{\cdot}16$, gives the largest value of Θ and is concave upwards whilst the most stable band $0{\cdot}01 < s < 0{\cdot}017$ has the smallest values and is concave downwards. Fig. 6 gives curves for the more stable data ($s > 0{\cdot}017$) classified in nine wind-speed bands which, allowing for the rather low statistical significance of the two extreme wind-speed bands, show a general tendency for Θ to increase for decreasing wind-speed and to be concave upwards for low U and concave downwards for high U.

7. Application to diffusion prediction

These curves have a direct bearing on the problem of estimating the diffusion of particles released continuously for a duration τ when sampled at a distance x downwind in a wind U. Hay and Pasquill (1959) gave the cross-wind width of such a plume (expressed

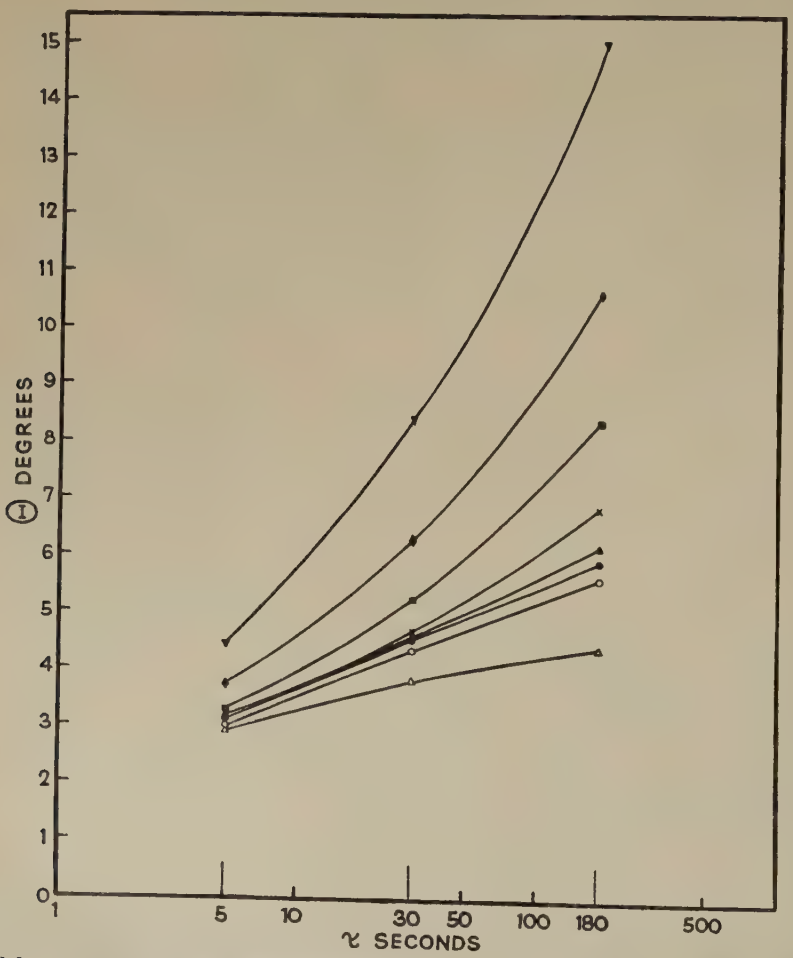
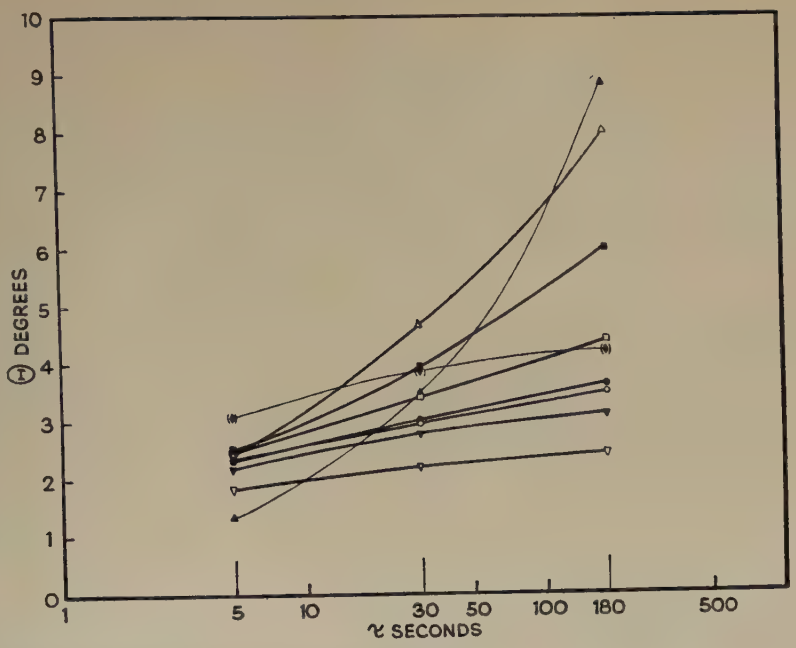


Figure 5. Mean values of Θ plotted against the sampling time τ for different stability ranges below $s = 0{\cdot}017$, when Θ is approximately independent of U, except through s.

$\triangle$ $+0{\cdot}017 \geqslant s \geqslant +0{\cdot}01$
$\bigcirc$ $+0{\cdot}01 \geqslant s \geqslant 0$
$\bullet$ $0 \geqslant s \geqslant -0{\cdot}01$
$\blacktriangle$ $-0{\cdot}01 \geqslant s \geqslant -0{\cdot}02$
$\times$ $-0{\cdot}02 \geqslant s \geqslant -0{\cdot}04$
$\blacksquare$ $-0{\cdot}04 \geqslant s \geqslant -0{\cdot}08$
$\blacklozenge$ $-0{\cdot}08 \geqslant s \geqslant -0{\cdot}16$
$\triangledown$ $s < -0{\cdot}16$



gure 6. Mean values of Θ plotted against the sampling time τ for different wind-speed bands in m sec^{-1} when $s > 0{\cdot}017$.

▲ = 0-1 □ = 3-4 ▼ = 6-7
△ = 1-2 ● = 4-5 ▽ = 7-8
■ = 2-3 ○ = 5-6 ◆ = 8-9

(Symbols bracketed when there are less than 5 cases)

an angular standard deviation Θ_p subtended at the source) by a formula which can be ritten as

$$\Theta_p = \sqrt{[\Theta_\tau^2 - \Theta^2_{x/4U}]}$$

The value 4 in the last term is an empirically determined ratio of the Lagrangian nd Eulerian time-scales. Thus, Fig. 2 represents the angular standard deviation of the ume when sampled for three minutes at a distance $x = 20\ U$ metres downwind. How-ver, a note of caution must be sounded when applying these results in very stable condi-ons. In a preliminary analysis of diffusion studies at O'Neill, U.S.A., Barad (1959) as reported that the ratio of the time-scales decreases, in some cases to less than one, as ie mean wind speed decreases and as the stability increases, inferring smaller diffusion han is implied by the above equation and Fig. 2.

Acknowledgments

The sensitive recording wind-vane used in these investigations was designed by Mr. J. I. P. Jones, C.D.E.E., Porton. The authors are indebted to the Director-General, Meteorological Office, for permission to publish this paper.

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551.510.41 : 551.510.534 : 551.521.3

Radiative transfer in the lower stratosphere due to the 9·6 micron band of ozone

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(Manuscript received 31 January 1961)

SUMMARY

Precise numerical calculations of the flux divergence due to narrow regions of the ozone 9·6 micron band were made, using the relative strengths and positions of the spectral lines calculated by Kaplan, Migeotte and Neven (1956). The absolute line strength was estimated on the basis of Walshaw's (1957) laboratory measurements. Spectral models or Curtis's approximation were not used, but Lorentz shapes and uniform widths were assumed for all lines. Calculations up to a height of 33 km were made for two ozone ascents of Brewer and Milford (1960). The atmospheric heating rates obtained for narrow regions of the band were then combined to estimate the heating rate by the whole band with results in qualitative agreement with, though nearly twice as great as those of Plass (1956a). The effect of the radiative temperature of the atmosphere below the ozone layer (whether of ground or cloud top) appears to be great. The radiative heating by ozone in the 10-20 km region is 0·2 or 0·3 degrees C per day for a ground temperature of 10°C, but may disappear altogether, if cloud is present just below the tropopause.

Using a larger computer, the procedure could be extended to cover the whole band, thereby avoiding the approximate extrapolation employed in this paper. The procedure could be applied to any absorption band, and is flexible to permit the use of any desired pressure and temperature dependence of the shape, width and strength of the spectral lines; results from it might serve as a standard against which conclusions reached by approximate methods can be checked.

1. INTRODUCTION

The three gases which contribute in important ways to the radiation budget of the atmosphere are water vapour, carbon dioxide and ozone. In the troposphere only the first two are important; the distribution of the radiative fluxes due to them may be calculated by various radiation charts, for example those of Elsasser (1942) or Yamamoto (1952). The accuracy of these is borne out reasonably well in actual measurements where these have been attempted (see Brewer and Houghton 1956).

In the lower stratosphere, the divergence of the radiative fluxes is much smaller, and it is not possible to obtain sufficient accuracy using radiation charts. Dobson, Brewer and Cwilong (1946) suggested that the three gases, water vapour, carbon dioxide and ozone, contributed similar amounts; the water vapour and carbon-dioxide bands probably resulting in cooling of this region and the ozone band in the heating of it. More precise calculations have been attempted since then for carbon dioxide and ozone, though none, as yet, for water vapour.

The main difficulty in making accurate computations of radiative transfer in the atmosphere comes from the fact that, in general, one is concerned with paths along which pressure, temperature and concentration of absorber all vary continuously. Various approximate methods have been developed for dealing with this problem. These are more applicable to calculations in the carbon-dioxide bands, which consist of regularly-spaced strong lines, and are made easier by the fact that the mixing-ratio of carbon dioxide is nearly constant with height. Such computations for the carbon-dioxide bands have been made by Plass (1956b) and Brooks (1958).

Calculations on the 9·6 μ ozone band have been carried out by Plass (1956a) who used the early laboratory absorption measurements of Summerfield (see Strong 1941). More recent measurements by Walshaw (1957) have, however, thrown doubt on the accuracy of the earlier data, and recent calculations of the positions and intensities of lines in part of the ozone band by Kaplan, Migeotte and Neven (1956), have made it clear that the complexity of the ozone band is such, that the approximate methods for solving the transfer

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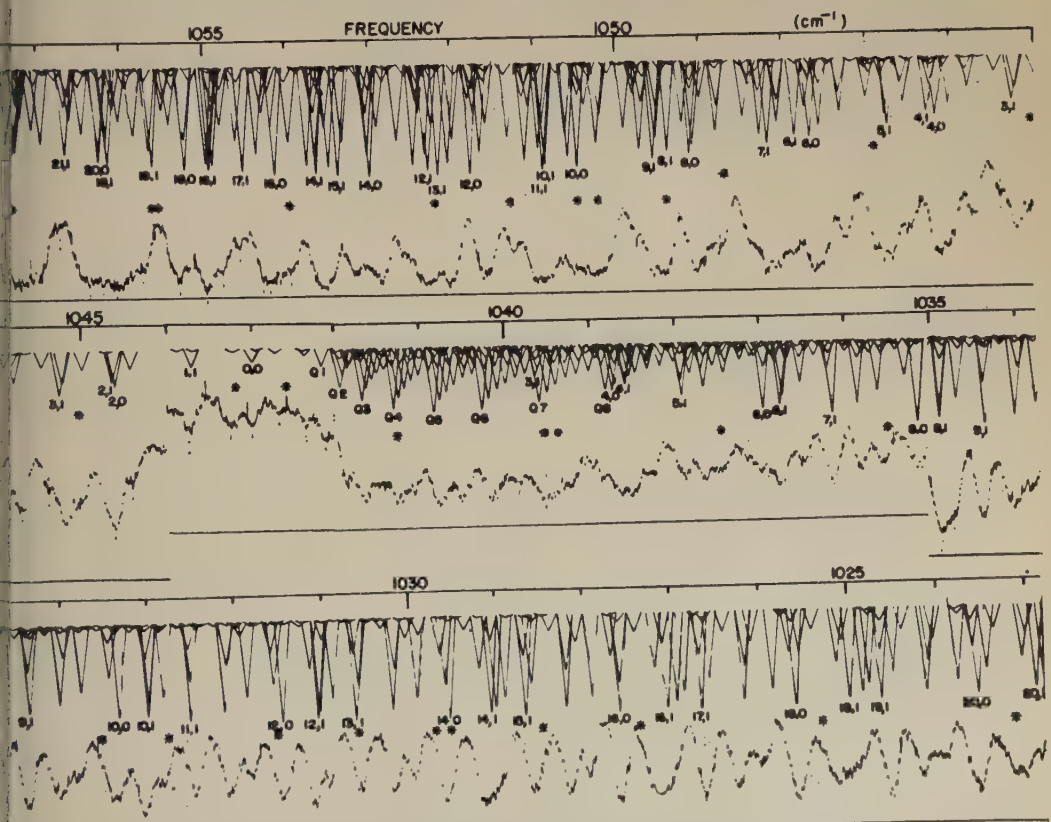
blem used in the case of carbon dioxide cannot be used with any confidence for the ne band. Because this detailed spectroscopic information is now available, it has come desirable to calculate radiative transfer through the 9·6 μ ozone band by direct merical integration of the transfer equations, without the use of doubtful models. re-examination of the ozone contribution to the radiative transfer in the stratosphere is o indicated in view of the more precise data of the ozone distribution obtained with emical sondes by Brewer and Milford (1960). The work here reported is based on ch ascents, which normally reach 27 km, and account for 70 per cent of the total ozone the atmosphere; extrapolation to greater heights has been made by reference to typical *kehr* measurements.

We have confined our computations of radiative divergence to the region below km. We did not, in this first attempt, go higher, since owing to Doppler broadening, e line-shape formulae we used are not valid above this height. Also, it is in the lower atosphere, that the divergence of long-wave radiation, though small, is competitive th the small divergence of solar radiation; hence accurate computations are required re to determine the magnitude and the sign of the overall radiative effect. It is also re that radiation processes relevant to the problems of the tropopause should become parent.

As will be seen, the method we describe is of more general application. It is an easy atter in our scheme to vary line-shape, line-width and line-strength in any desired anner with pressure and temperature. It may thus be applied to calculate radiative ansfer through any absorption band without any simplifying assumptions, providing at adequate spectroscopic information is available.



igure 1. The part of the 9·6 μ band of ozone calculated by Kaplan, Migeotte and Neven (1956). The lines re represented by triangles, and heights are indicative of line strengths. The lower curve is part of the solar absorption spectrum obtained at the Jungfraujoch.

2. Radiative transfer by the 9·6 μ band of ozone

A detailed study of the structure of part of this band has been made by Kaplan, Migeotte and Neven (1956). Using the model of an asymmetric top for the ozone molecule they have calculated the positions and relative line strengths of lines in the band from 1,023 cm^{-1} to 1,058 cm^{-1}. Fig. 1, which is adapted from their original paper, shows the calculated positions and relative line strengths compared with observations of the solar spectrum made from the Jungfraujoch. The complicated structure and the very close spacing of the lines of the band is immediately obvious. Accurate comparison between the calculated and observed spectra is difficult, because the resolution of the observed spectra, although better than 1 cm^{-1}, is nothing like sufficient to resolve all the detail. However, agreement between the observed and calculated spectra seems satisfactory near the centre of the band, but falls off at the extremes for which calculations have been made. Kaplan *et al.* explain that this is either due to centrifugal stretching of the molecule at higher rotational quantum numbers, or to a Coriolis interaction, neither of which they are able accurately to allow for at the moment.

A thorough laboratory study of the absorption in this band for different amounts of ozone and different pressures has been made by Walshaw (1957). The resolution of his instrument was only about 6 cm^{-1}, and so his measurements cannot be used to compare with calculated individual line positions and intensities. We have used them, however, as explained in Section 3, to obtain a factor which will convert the relative line strengths given by Kaplan *et al.* to absolute line strengths.

The shape of an absorption line under the conditions we are considering is almost entirely determined by collision broadening, for which the absorption coefficient k_ν at a frequency ν can be described by the Lorentz equation

$$k_\nu = \frac{S\gamma}{\pi\left[(\nu - \nu_0)^2 + \gamma^2\right]} \quad , \qquad . \qquad . \qquad . \qquad (1)$$

where S is the line strength, ν_0 is the frequency of the line centre and γ is the half-line width, which is directly proportional to the collision frequency in the gas and so may be written

$$\gamma(p, T) = \gamma_0 \frac{p}{p_0}\left(\frac{T_0}{T}\right)^{\frac{1}{2}} \qquad . \qquad . \qquad . \qquad . \qquad (2)$$

Here γ and γ_0 are the values of the half-line widths at pressures p and p_0 and temperatures T and T_0 respectively. The variation of p in the atmosphere is clearly very much greater than that of $T^{\frac{1}{2}}$; in our calculations a mean value for T has been assumed and its variation neglected. At pressures less than about 7 mb the Doppler effect (see, for instance, Goody 1958) seriously modifies the line shape, but at heights below about 33 km this may be neglected.

For calculations on the absorption of infra-red bands it has generally been assumed that γ_0 has the same value for all the lines in the band. This is almost certainly not true (see, for instance, Benedict and Kaplan 1959); its value may well vary by as much as a factor of 3 for lines in the same band, but no detailed or useable information is at present available. However, since γ is subject to a 100-fold variation due to changes in pressure, it is felt that the introduction of the much smaller variation of γ_0 for different lines would not greatly alter our results. Accordingly, we took the mean value of γ_0 as given by Walshaw (1955) to be 0·081 cm^{-1}; for a mean value of 240°K for the temperature, this becomes 0·086 at 1,000 mb. It should also be pointed out that the strength S of the line varies with temperature because the populations of the energy levels are a function of temperature. We have not taken this into account and have used a mean temperature of 240°K. The temperature range in the atmosphere over the levels studied is not very great, and neglect of this factor will make little difference to our results. Since S, the strength of the line and m, the mass of absorber enter into the equations in exactly the same way, neglect of the effect of temperature on S will be equivalent to small errors in the ozone concentration.

Now the transmission τ_ν at a given frequency ν of a path of atmosphere of length x which the density of absorber is ρ is given by

$$\tau_\nu = \exp(-k_\nu \rho x). \quad . \quad . \quad . \quad . \quad (3)$$

s relation, known variously as Beer's or Bouguer-Lambert's absorption law, is always d in this monochromatic form, and serves to define k_ν. The product ρx can be replaced the single symbol m, which is the mass of absorber in a column of unit section with its s along the direction of propagation of the radiation. Frequently m is expressed in ms of the height this column of absorber would have at S.T.P. Possible units for m thus g cm^{-2} or cm (S.T.P.).

Eq. (3) assumes that k_ν is constant along the path. In general, it is not, but is a function of pressure and temperature, and we must write

$$\tau_\nu = \exp(-\int k_\nu \, dm),$$

$$\tau_\nu = \exp(-\sum_i k_{\nu i} m_i). \quad . \quad . \quad . \quad . \quad (4)$$

e second form implies that the layer is divided into thin slices, each with an effective sorption coefficient of $k_{\nu i}$ and containing an absorbing mass m_i; it is thus suitable for merical computation.

There are two approximate combinations of Eqs. (1), (2) and (4), which enable the tegration over frequency to be carried out analytically and which have been widely ed in studies of the radiative transfer in the atmosphere. They have been studied tensively by a number of workers, in particular Plass (1954 and 1960) and Godson 955). The first of these applies when the absorption is small and the exponential may represented by the first term of its expansion. The integrated absorption of a line is en equal to Sm and is independent of p. The other approximation arises when the sorption due to a line at the line centre is so strong that the transmission there is very hall indeed; the line is then said to be ' black,' and k_ν may then be written

$$k_\nu = \frac{S\gamma}{\pi(\nu - \nu_0)^2} \quad . \quad . \quad . \quad . \quad (5)$$

his approximate value of k_ν is the same as in Eq. (1), if $(\nu - \nu_0) \gg \gamma$, while near the line ntre the value of k_ν is much larger; however, since the transmission there is very small any case, this is not of practical importance. Using Eq. (5), Eq. (4) appears in the rm

$$\tau_\nu = \exp\left(-\frac{S\gamma_0}{\pi p_0 (\nu - \nu_0)^2} \int p\,dm\right) \quad . \quad . \quad . \quad (6)$$

here Eq. (2) has been used for γ and the variation with temperature neglected. Eq. (6) an be further simplified by the use of a mean pressure $\bar{p}$ defined for the layer by

$$\bar{p} = \frac{1}{m} \int p\,dm \quad . \quad . \quad . \quad . \quad (7)$$

he resulting form for τ_ν may then be amenable to an analytical integration over frequency.

The use of the mean pressure defined by Eq. (7) was first suggested by Curtis (1952); is clearly valid providing either of the above approximations may be used. That the econd approximation is inapplicable in the case of ozone has been pointed out by Goody 1957) and Kaplan (1957), because the lines in this band are not strong enough to make he approximation valid for the ozone concentrations found in the atmosphere. That either approximation is applicable to ozone is demonstrated in Fig. 2, where curves f τ_ν for several actual paths in the atmosphere are presented; the lines are, in general, too trong for the first approximation to apply, and not strong enough for the second. It is also

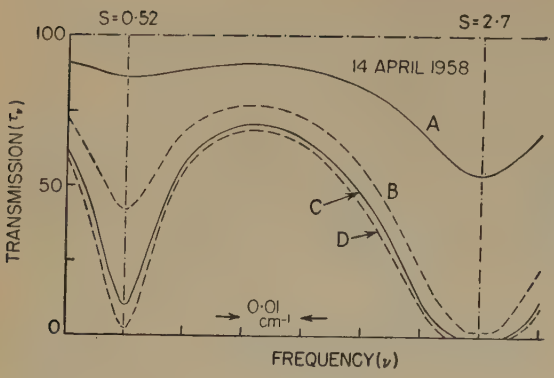


Figure 2. Monochromatic transmission vertically through four layers of atmosphere; 10-15 km (A), 10-20 km (B), 10-25 km (C), 10-30 km (D), in the vicinity of two spectral lines in the 9·6 μ band. Ozone distribution as in Fig. 8. The lines are spaced 0·06 cm^{-1} apart and have strengths 0·52 and 2·7 on the relative scale used by Kaplan *et al.* (1956).

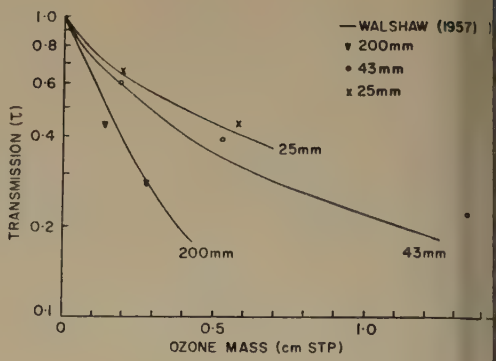


Figure 3. Transmission through ozone in the spectral interval 1026 to 1028 cm^{-1} at three pressures (in mm of mercury). The curves are plotted from the measured data of Walshaw (1957) and deposited by him with the Royal Meteorological Society. The marked points were computed by the method of section 3 from Kaplan's calculated line strengths. To obtain the agreement indicated in the figure, the theoretical relative line strength values were multiplied by 0·2.

quite clear from Walshaw's (1957) measurements (see also, Fig. 3) that it is not permissible to describe the transmission through ozone in these quantities and at these pressures either by a single function of m or of pm. Because of this, we have carried out a direct numerical solution of the radiative transfer equations for selected parts of the ozone band. These calculations will now be described.

The complete theory of radiative transfer permits the calculation of the flux at any height z originating in the atmosphere both above and below z. A complete statement of the upward flux at height z is (see for instance Elsasser, 1942)

$$F_{\uparrow} = \int_{\theta=0}^{\pi/2} \sin 2\theta \int_{z'=z}^{0} \int_{\nu=\nu_1}^{\nu_2} B_{\nu}(z') \frac{d\tau_{\nu}(z, z')}{dz'} d\nu \, dz' \, d\theta \quad . \quad . \quad (8)$$

where $B_{\nu}(z')\, d\nu$ is the energy emitted by unit area of a black-body surface at the temperature corresponding to the height z' between the frequency limits ν and $\nu + d\nu$, $\tau_{\nu}(z, z')$ is the transmission at frequency ν of the atmosphere between levels z' and z measured along a path making an angle θ with the vertical, and it is calculated according to Eq. (4) with $dm = \rho \sec\theta \, dz''$, the variable z'' being taken from z' to z. The limits ν_1 and ν_2 correspond to the effective width of the band. For the flux from above $F_{\downarrow}$, the same equation holds except that the limits of integration over z' should be changed from $(z, 0)$ to (z, ∞).

The net flux F is given by

$$F = F_{\uparrow} - F_{\downarrow} \quad . \quad . \quad . \quad . \quad . \quad (9)$$

The divergence of F is generally written as a 'heating rate' dT/dt which is the rate at which the air temperature at height z *tends* to be modified by the supply or loss of radiational energy. Thus

$$\left(\frac{dT}{dt}\right)_{\text{radiation}} = -\frac{1}{c_p \delta}\left(\frac{\partial F}{\partial z}\right) \quad . \quad . \quad . \quad (10)$$

where δ is the density of the air at height z and c_p is its specific heat at constant pressure.

It is evident that the detailed evaluation of the fluxes, and hence of the heating rate, a problem of great complexity, which in general, requires simplification. The most dical simplification would result if the integration over frequency could be performed st, once and for all. This is precisely where the formulation of suitable spectroscopic dels checked against laboratory measurements play an important part. But, as has en explained here, methods for doing this in the case of ozone are not now available. more modest possibility of simplifying expression (8) depends on finding a short-cut the integration over θ. It has been suggested that for water vapour and carbon dioxide good approximation results from the assumption that sec θ can be set equal to a constant both 1·5 and 1·66 have been used; e.g., Roberts 1930; Kaplan 1952). We have subjected is suggestion to a careful test, and found the factor 1·667 appropriate for ozone.

To do this, the integrals over z and ν in expression (8) which we may call $I_\uparrow(z, \theta)$ $I_\downarrow(z, \theta)$ were evaluated by the methods to be described in Section 3 for three different lues of θ (25°, 53° and 72½°), so chosen that the corresponding values of sec θ are in the nvenient ratios 1 : 1·5 : 3 and with sec 53° = 1·667. (These calculations were made r the ascent of 2 May and for the spectral interval near 1051 cm^{-1}). The remaining tegration over θ was then carried out numerically using the form

$$F_\uparrow(z) = \int_0^{\pi/2} I_\uparrow(z, \theta)\, d\left[\tfrac{1}{2}(1 - \cos 2\theta)\right] \quad , \qquad . \qquad . \qquad (11)$$

d with a similar expression for $F_\downarrow(z)$. The values of $I(z, \theta)$ for different values of θ ere fitted to a cubic equation in the variable x where $x = \frac{1}{2}(1 - \cos 2\theta)$, with the rther condition that $dI/dx = 0$ at $x = 0$. Integral (11) could then be evaluated analy- cally, and the flux divergence at all heights could be deduced. The resulting heating ites differed at all levels by less than 5 per cent and for most levels by less than 1 per cent om what would have been obtained if we had put $\theta = 53°$ (i.e., sec $\theta = 1{\cdot}667$), and mitted the integrations over θ.

A further simplification may be made to expression (8). Over the entire absorption and (say from 9 to 10 μ), B_ν varies smoothly and by less than 6 per cent. One is thus stified in assuming a mean value for B_ν, say B, and removing it from the integration ver frequency. Using both these simplifications, (8) may be written :

$$F_\uparrow = \int_z^0 B(z') \int_{\nu_1}^{\nu_2} \frac{d\tau_\nu(z, z')}{dz'}\, d\nu\, dz', \qquad . \qquad . \qquad . \qquad (12)$$

with

$$\tau_\nu(z, z') = \exp\left[-\int_{z''=z'}^{z} 1{\cdot}667\, k_\nu\, dm(z'')\right] \qquad . \qquad . \qquad (13)$$

r in a form more suitable for numerical computation :

$$F_\uparrow = \sum_{z'=z}^{0} B(z')\left[\Delta\left\{\sum_{\nu_1}^{\nu_2} \tau_\nu(z, z')\, \Delta\nu\right\}\right] \qquad . \qquad . \qquad . \qquad (14)$$

with

$$\tau_\nu(z, z') = \exp\left\{-\sum_{z''=z'}^{z} 1{\cdot}667\, k_{\nu i}\, m_i(z'')\right\} \quad . \qquad . \qquad . \qquad (15)$$

A similar expression is obtained for $F_\downarrow$, the only difference being that the upper limit or z' is now ∞.

3. The calculations

Even in the simplified form of Eq. (14), flux (and hence, heating rate) calculations are a large undertaking, since the 9·6 μ band of ozone extends over more than 100 wave-numbers, and contains about 500 lines. The precision required is high because each calculation involves two differentations – one in the fluxes ($\partial\tau_\nu/\partial z$ in Eq. (12)), another

in the flux divergence (Eq. 10). Consequently, a large number of ordinates have to be calculated for the integrations, especially for that over ν, since the variation of τ_ν with ν is quite rapid near the line centres; in fact, something like 20,000 values of τ_ν over the whole band would be required to give adequate precision. Whilst this is feasible in a reasonable time with the use of computing machines of the highest speed, we could not attempt the whole problem on the Ferranti Mercury of the Oxford University Computing Laboratory. Accordingly, we studied the heating rates of three relatively narrow parts of the band and of single lines of eight different strengths. We then combined these partial heating rates to obtain that produced by the whole band. Comments on the machine time required are made in the Appendix.

From the part of the 9·6 μ band analysed by Kaplan *et al.* (1956) we selected for detailed analysis three spectral regions listed in Table 1. These differ considerably in average line spacing and line strength, but reflect well the extremes revealed by their tabulation of Kaplan *et al.**. The fourth and bigger sample in the Table was selected for calibration only, a procedure intended to relate Kaplan's relative line strengths to absolute line strengths.

TABLE 1. SAMPLES OF THE OZONE 9·6 μ SPECTRUM SELECTED FOR DETAILED ANALYSIS, AND FOR CALIBRATION

Location cm^{-1}	Limits of sample cm^{-1}	Width of sample cm^{-1}	No. of lines in sample	Avg. line spacing cm^{-1}	Line strengths† in sample		
					Min.	Mean	Max.
1029	1028·995 1029·535	0·54	6	0·09	0·1	5	10
1041	1040·995 1041·235	0·34	9	0·038	0·5	2·5	5·1
1051	1051·655 1052·075	0·42	8	0·05	0·3	4	8·8
1027‡	1025·995 1027·905	1·91	20	0·095	0·4	4·4	13·7

† line strengths as listed in Kaplan, Migeotte and Neven (1956)
‡ this sample selected for calibration only

In addition, single lines of strengths 0·173, 0·52, 0·9, 1·56, 2·7, 4·68, 8·1 and 14·08 were studied. For each of these, the spectral intervals considered were : line centre $\pm$ 0·065 cm^{-1}, and line centre $\pm$ 0·245 cm^{-1}.

(*a*) *Calibration*

This was done by comparing the transmission for two wave-numbers centred at 1,027 cm^{-1} and calculated by direct numerical integration using line strengths and positions from the Kaplan *et al.* Table with Walshaw's laboratory measurements. Fig. 3 shows Walshaw's curves, and our eight calculated points, adjusted for a calibration factor (to be applied to Kaplan *et al*'s. line strengths) of 0·2. The mean deviation from this mean value in the eight cases is less than 0·05. The fact that agreement is not uniformly good is probably related to the great difference in the spectral resolution : 2 cm^{-1} for the computation and 6·5 cm^{-1} for the measurements. The agreement is, however, quite close for the smaller ozone masses of meteorological interest.

The integrations under the resolved spectrum were performed as follows; the same procedure was used in the flux calculations. This sample interval of 2 cm^{-1} contained 20 lines. The absorption coefficient at any frequency within this interval depends, of course, not only on these 20 lines, but also on lines outside the interval. We found it necessary to consider 60 such exterior lines, but it turned out to be possible with very small error to bunch these 60 lines into 25 groups located up to 10 cm^{-1} away from the sample. Integrations (of the form $\int_\nu [\exp(-k_\nu m)]\, d\nu$) were performed separately near

* Deposited as Document 4,853 with the Photoduplicating Service of the Library of Congress, Washington 25, D.C.

e line centres, and in the smoother (and therefore, less exacting) regions between the les. We considered each of the line centre regions to be 0·01 cm^{-1} in width, centred the line. Each such interval was divided into 5 strips of width 0·002 cm^{-1}, and in ch of these we employed a Gauss five-point integration formula. This means that the sorption coefficient was calculated precisely at 25 points for each absorption line, at an *erage** spacing of 0·0004 cm^{-1}. For the integrations between the lines, five-point lynomial formulae were employed, calling for calculations of the absorption coefficient intervals 0·0025 cm^{-1}. The relative weights attached to successive ordinates were 32, 12, 32 and 7.

These particular integration procedures were chosen on the basis of a preliminary udy in which integrals obtained from ordinates of widely different spacings were compared. ear the line centres, one test established the superiority of the Gauss formula over the lynomial type when similar numbers of ordinates are used. In another test, a single ne of strength 2 was studied over wide ranges of pressure (8 to 100 mb) and ozone mass ·1 to 0·01 cm s.t.p.), so that in some cases the line was strongly saturated and produced arp discontinuities. Results obtained from formulae using 15 and 25 ordinates per line ffered by less than 0·01 per cent, those from 25 and 45 ordinates differed by less than 0001 per cent. It is believed, therefore, that the adopted 25-ordinate procedure is more an adequate. It is quite probable that an integration based on 15 ordinates per line ould have sufficed. To justify our procedures for the integrations between lines, we tegrated the interval 0·005 to 0·065 cm^{-1} from the line centre. The spacing of the rdinates in the interval was varied in seven steps from 0·015 to 0·000833 cm^{-1}. Pressure hd ozone mass were allowed to have values as in the other tests. An orderly convergence f the resulting integrals, corresponding to finer and finer ordinate spacings, was noted. was concluded that a spacing of 0·004 cm^{-1} was probably adequate, and that the actually dopted spacing of 0·0025 cm^{-1} produced integrals differing from those obtained with ne finest spacing by less than 0·01 per cent in all cases, and by much less at pressures reater than 10 mb.

(*b*) *The ascents*

The ozone and temperature information required was taken from the Liverpool oundings reported by Brewer and Milford (1960) for 14 April and 2 May 1958, and is lisplayed in Figs. 8 and 11. In addition we considered a somewhat artificial modification f the latter ascent, in which some of the ozone above 9 km was shifted into the troposphere, nd given a distribution there similar to the one that occurred over Liverpool on 20 April 958. At the same time, a coarser grid was employed in the calculations, the atmosphere eing divided into layers 1·25 km thick, in place of the 1-km layers otherwise used. The urpose of this modified sounding was to study the effect of low-level ozone and to observe whether the coarsening of the grid seriously affected the heating rate results. All soundings gave reliable information to heights close to 27 km. Beyond this level, the ozone distribu- ion was extrapolated on the basis of the total-ozone information available from Professor Dobson's ozone spectrograph at Oxford and in conformity with the *Umkehr* measurements collected by Götz (1951). The profiles thus extend from the ground into that region of he middle stratosphere, where the temperature rapidly increases with height. It includes of course the tropopause and the thick nearly isothermal layer above it in which the ozone reaches its maximum (absolute) concentration.

(*c*) *The programmes*

The radiation fluxes and divergences were calculated for three sample regions of the spectrum and for single absorption lines, as described in Table 1. The direct integration under the absorption curves in each of the samples required about 330 values of the mono- chromatic transmission. The first step consisted in building up in the store of the computer,

* The Gauss integration formulae require non-uniformly spaced ordinates. The relevant formulae are listed in texts on numerical methods. Hildebrand's (1956) discussion was found especially helpful.

values of the absorption coefficient at 330 values of the frequency and appropriate to the temperature, pressure and ozone mass in 25 layers (index i) in the atmosphere. The resulting 8250 numbers constitute a matrix of 25 rows and 330 columns, which nearly filled the available slow-speed store. Each element of the matrix is of the form

$$a(\nu, i) = \sum_j \frac{1{\cdot}667\, S_j\, m_i\, \gamma_i}{\pi\,[(\nu - \nu_j)^2 + \gamma_i^2]} \quad . \quad . \quad . \quad (16$$

where the summation is over the n lines which contribute to absorption at the frequency ν; n was about 40 in our cases. The half-line width

$$\gamma_i = \frac{\gamma_0\, p_i}{p_0},$$

p_i being mean pressure of the ith layer. Each element of the matrix is thus one term of the summation of Eq. (15) involved in finding values of τ_ν.

The depth of each of the 25 layers was 1 km for the two ascents of 14 April and 2 May 1958, and their mean heights were at 9·5, 10·5 etc., to 32·5 km. In addition, one layer represented the atmosphere above 33 km to account for the effect of the ozone beyond the range of these calculations. The contribution of the atmosphere below 9 km could be disregarded because of the negligible ozone concentration there. The ground radiates as a black body, which can, for mathematical purposes, be considered as a layer with an infinite ozone mass, and thus of zero transmission at ground temperature. If clouds are present these take the place of the ground, and their effect can be accounted for by setting the ' ground ' temperature equal to that of the cloud top. In all our calculations, we considered black-body temperatures below the ozone layers of 283, 263, 243 and 223°K. Thus our results include the effect of cloud tops with temperatures down to − 50°C. For the third (synthetic) ascent, appreciable ozone concentrations were assumed down to 2·5 km. Above this height, we considered 24 layers of depth 1·25 km, extending up to 32·5 km as before.

The second step in the calculation involved summations along columns of matrix elements : $A(z, z') = \sum_i a(\nu, i)$ corresponding to the summations at constant frequency of Eq. (15), the index i being taken over the layers between the heights z and z'. Here we allowed z' to range from 9·5 to 32·5 in steps of 1 km, with an additional value to represent the atmosphere above 33 km. The increments in i correspond to layers of 2 km thickness (2·5 km for the third ascent). The transmissions τ_ν are obtained as the exponentials of $(-A)$, as in Eq. (15); this is followed by summation over frequency by the methods previously discussed giving

$$\tau(z, z') = \sum_{\nu_1}^{\nu_2} \tau_\nu(z, z')\, \Delta\nu,$$

which is the inner summation of Eq. (14).

These values of τ were obtained for groups of layers 1 km thick (1·25 km for the modified ascent) from 10 km to 32 km (4·75 to 31·25 for the modified ascent); the upward and downward figures were found using Eq. (14) for levels 2 km apart (2·5 km for the modified ascent) from 11 to 31 km (5 to 30 for the modified ascent).

4. Results

As examples of the results obtained directly from the procedures outlined, Figs. 4, 5 and 6 show the contributions to the heating rates by the three spectral samples, and Fig. 7 shows the contribution by single lines of various strengths. All these curves apply to the

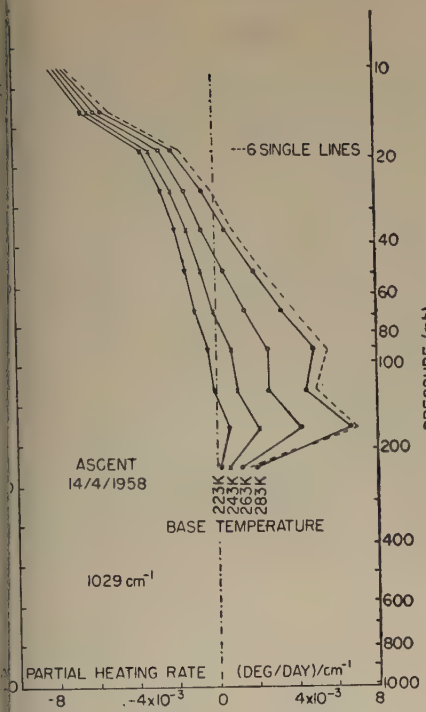


Figure 4. Sample at 1,029 cm^{-1}.

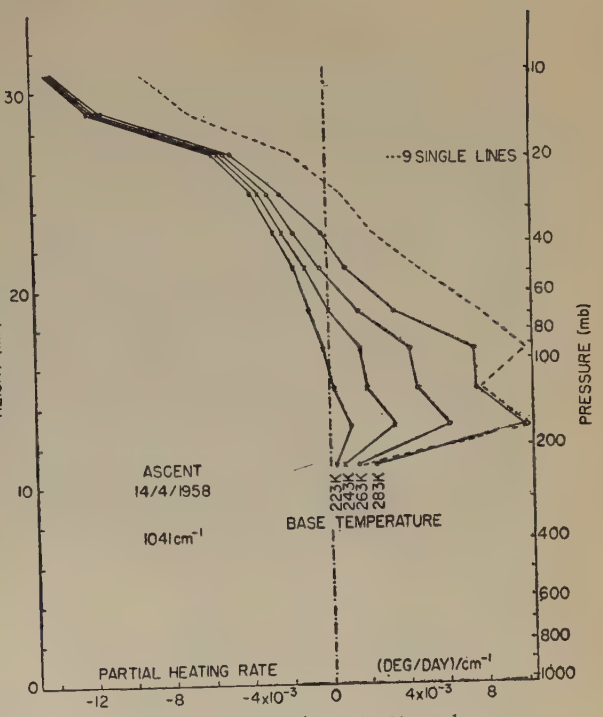


Figure 5. Sample at 1,041 cm^{-1}.

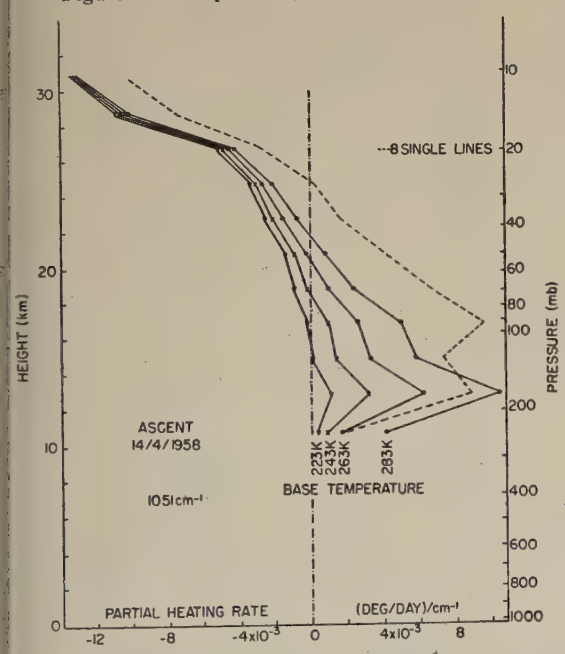


Figure 6. Sample at 1,051 cm^{-1}.

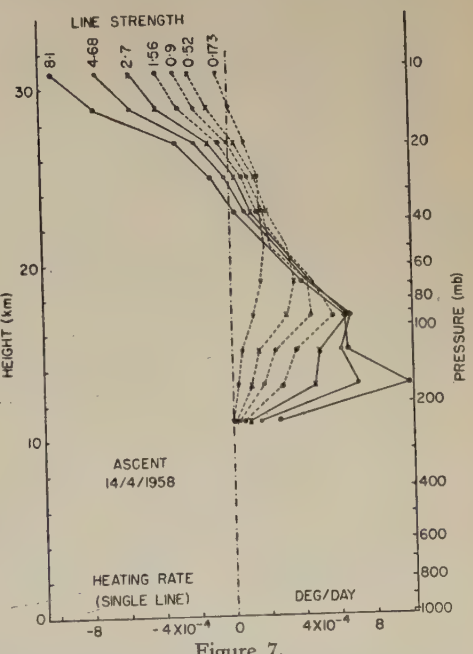


Figure 7.

Figures 4 to 6. Distribution with height of the heating rate in a spectral interval of one wave-number, calculated from the spectral samples of Table 1, and for the ascent of 14 April 1958. The four solid curves correspond to ground or cloud-top temperatures of : 10, — 10, — 30 and — 50°C. The dashed curve was constructed by adding the heating effects of single lines of about the strength occurring in the spectral interval.

Figure 7. Distribution with height of the heating rate by single lines whose strengths, on the relative scale of Kaplan *et al.* (1956) are indicated at top. Calculations are applicable to the 9·6 μ band and the ascent of Fig. 9, with a base temperature of 10°C. For each line, the heating rates are for a spectral interval : line centre $\pm$ 0·065 cm^{-1}, contributions of further outlying regions being negligible.

ascent of 14 April 1958 (Fig. 8). While there are differences in detail between the curves which in a qualitative way can be accounted for on the basis of the characteristics of the spectral samples and the line strengths, there are enough features in common to allow an attempt at combining the results to assess the contribution of the entire absorption band. All curves point to the importance of the temperature of the black body at the bottom of the atmosphere. A variation of a mere 20°C affects the resulting heating rate maxima in the 12 to 18 km interval by as much as a factor two. Such temperature changes can also shift the height of zero divergence by as much as 4 km. The temperature structure *above* the ozone layer is not nearly so important however, as is illustrated in Fig. 10, where we allowed for two additional effective temperatures of the ozone beyond the reach of the ascent. The resulting heating-rate curves show important differences only down to 29 km; no effect at all remained below 23 km.

An attempt was made to reproduce the partial heating rates of the three spectral sample regions by combining some of the single-line curves of Fig. 7. By adding flux divergences from six single lines of appropriate strengths, one of the divergence curves calculated for the 1029 cm^{-1} sample was obtained to a good approximation (Fig. 4). This is not surprising, since in this part of the band, the lines are relatively far apart (see Table 1). In the other two spectral regions, because of the considerable overlap of the lines, equally good agreement was not obtained.

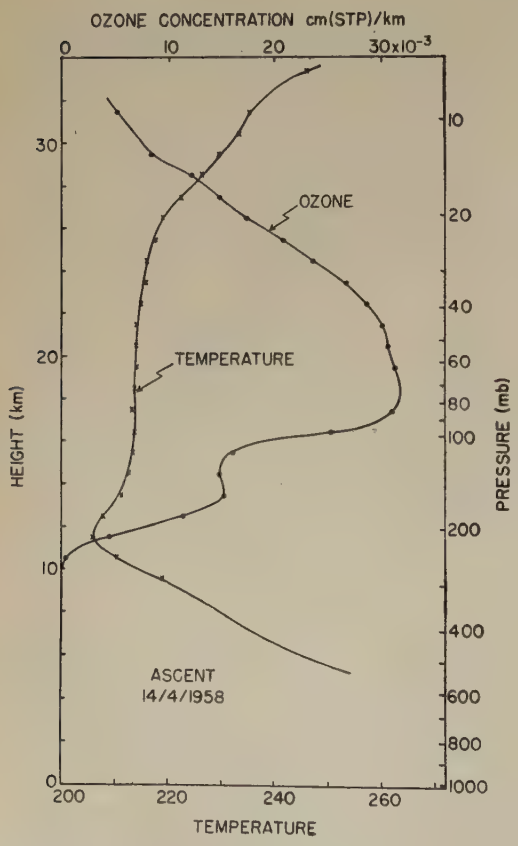


Figure 8. Temperature and ozone distribution over Liverpool on 14 April 1958. (From Brewer and Milford 1960). Total ozone over Oxford (from Dr. Dobson) 0·446 cm, ozone (assumed) above 32 km — 0·023 cm.

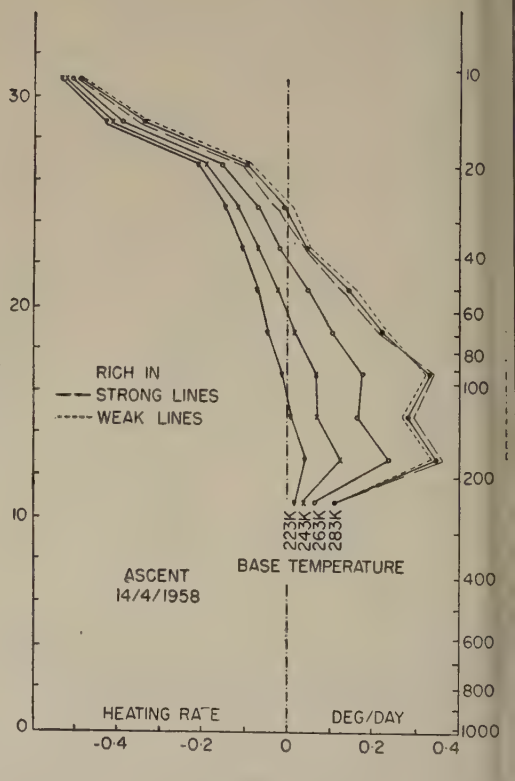


Figure 9. Heating rates due to the 9·6 μ band of ozone computed from the partial heating rates of Figs. 4 to 7 according to Table 2. The dashed lines indicate the small effect of varying the formula of Table 2, for a base temperature of 283 K.

Extrapolation to the whole band

To obtain the effect of the entire absorption band the observations of Migeotte, even and Swensson (1956) of the transmission of the entire atmosphere over the Jungaujoch were examined. This spectrum would suggest that in the atmosphere the significant extent of the band (absorption greater than 5 per cent) is 992 to 1,073 cm^{-1}. Of is, the interval 1,023 to 1,058 has been mapped by Kaplan *et al.* Starting in this latter gion, we drew up a distribution of spectral lines according to their strengths. This table as then extended to cover the entire band, using Migeotte's measured absorption as a uide. Such a distribution chart, in which the important central part is based on Kaplan's ata, allows one to divide the band into regions, approximately homogeneous in character. Where the overlapping of lines is serious, the nature of the spectrum was assumed to be rpified by one of the three samples we had selected. Thus, the 17 wave-numbers 1,021-,033, 1,046 and 1,061-1,063 are like the 1,029 cm^{-1} sample; and 8 wave-numbers 1,034-,041 like the 1,041 cm^{-1} sample; and the 14 wave-numbers 1,047 to 1,060 like the 1,051 m^{-1} sample. Beyond these regions, the lines are sufficiently sparse that the heating ates of individual lines can be added. The estimation of the entire band thus reduces to

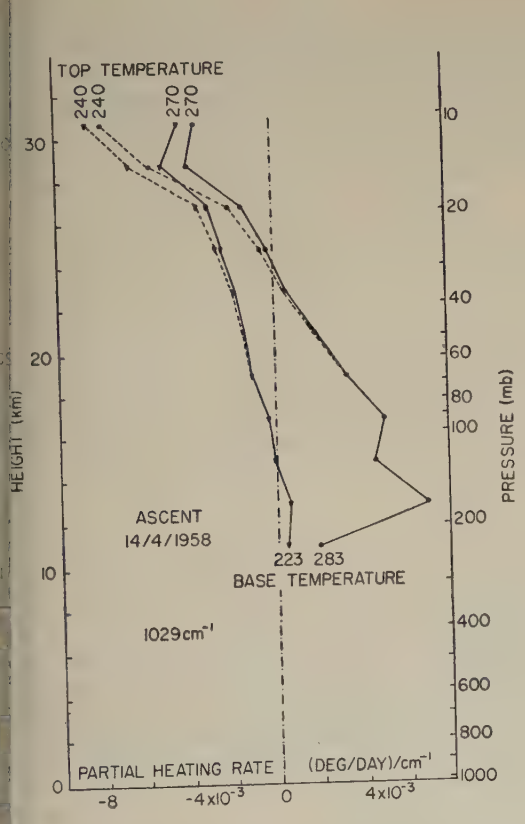


Figure 10. Showing the effect of varying the temperature of the layer above 33 km on two of the curves illustrated in Fig. 4. (For Figs. 4 to 7, the effective top temperature was taken to be 252°K, our best estimate).

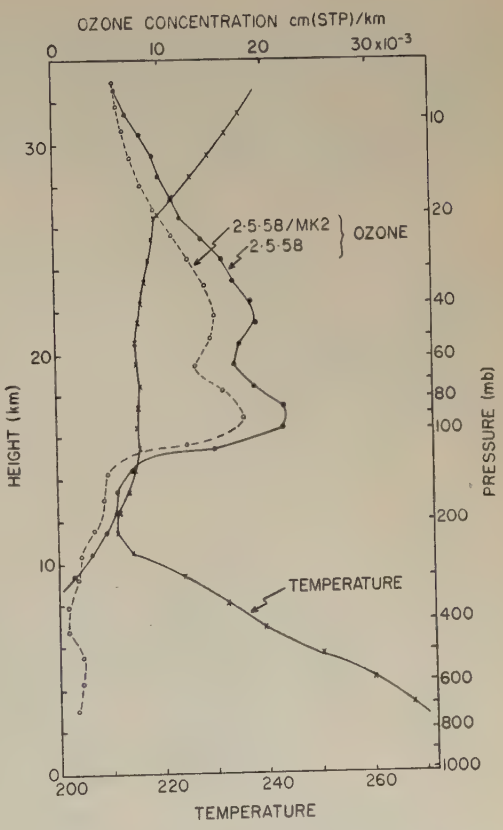


Figure 11. Temperature and ozone distribution on 2 May 1958 (see Fig. 9). The ozone distribution marked 2·5·58/MK2 is an artificial modification, in which some 4 per cent of the total ozone was assumed to be below 10 km, and distributed there according to the distribution actually observed on 20 April 1958. Total ozone over Oxford (from Dr. Dobson) 0·317 cm; ozone (assumed) above 32 km — 0·020 cm.

TABLE 2. HEATING RATE DUE TO THE OZONE 9·6 μ BAND

$= 17 \times$ heating rate of 1029 cm^{-1} sample

$+ \ 8 \times$ heating rate of 1041 cm^{-1} sample

$+ 14 \times$ heating rate of 1051 cm^{-1} sample

+ One of the combinations of the heating rates of single lines. Three combinations were tried:

Weak	Best	Strong	Heating rate of single line of strength
200	120	50	× 0·173
120	80	50	× 0·52
50	50	30	× 0·90
20	40	55	× 1·56
–	10	25	× 2·70
–	10	25	× 4·68

the sort of formula set out in Table 2. In the parts of the band not covered by Kaplan's calculations, some guess-work was necessary. But one can estimate the full extent of error here by deliberately exaggerating the number of strong or weak lines while still keeping the total absorption due to all the lines in agreement with the observations of Migeotte *et al.* The resulting formulae are also shown in the Table, and their effects can be seen directly in Fig. 9.

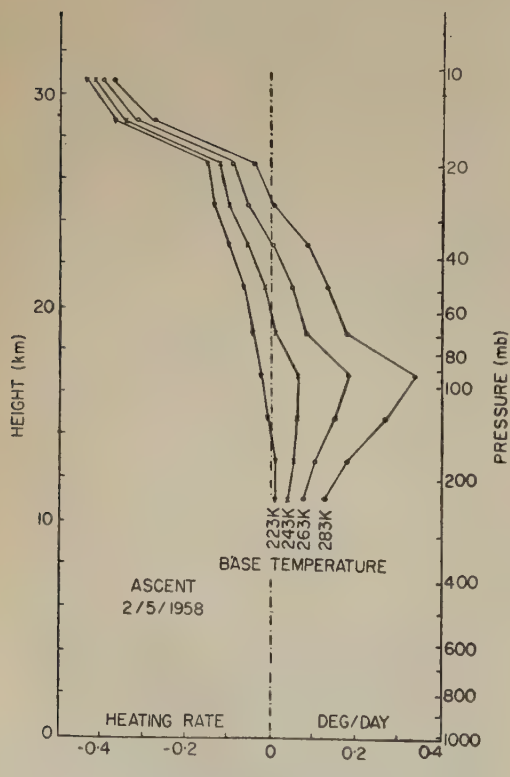


Figure 12. Heating rates for 2 May 1958 obtained by the same methods as Fig. 9.

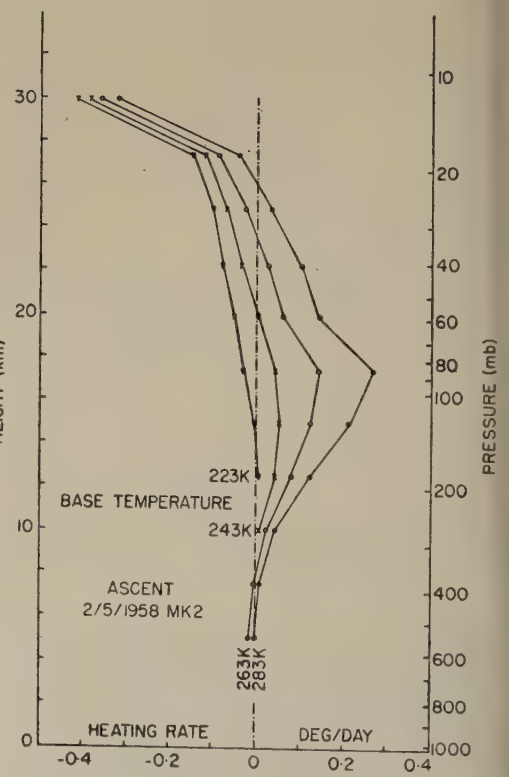


Figure 13. Heating rates for the ascent shown in Figure 11 as 2·5·58/MK2.

The final results obtained for the three ascents are displayed in three groups of figures. or the ascent of 14 April, Fig. 9 shows the heating rates for four base temperatures. ig. 8 and 11 show the ascent information). For one of the curves we have indicated e results of varying the extrapolation formula within the extremes indicated in Table 2. nce the effect is very small (here and in all other cases), the other curves show merely e results based on the 'best' formula. The two maxima in the ozone distribution rve of Fig. 8 show up quite markedly; the difference between them is somewhat reduced nce the major maximum at 18 km lies in the almost isothermal part of the stratosphere. he absolute values of these maxima (0·35°C/day) are higher than had previously een estimated, and might well be dynamically significant. In the presence of cloud, the eating rates are very much less at all levels, and it is only at considerable heights (25 km) hat the specific effect of the ground temperature disappears. For this same reason, there no single height in the atmosphere at which the isothermal layer above the tropopause educes the heating rate to zero. It would seem that this can happen anywhere between 4 and 24 km. The cooling rates above these levels are numerically greater than the heating ates below, but the curves cannot be expected to diverge much further with increasing eight, and must return to the zero axis below the mesopeak (50 km).

Figs. 12 and 13 show the results for the other two ascents. It would appear that the oarsening of the grid from 1 to 1·25 km does not wipe out the significant features, and rom the radiation point of view the effects of the small ozone concentration in the tropo- phere are probably not significant. The ascent showed two ozone maxima at 17 and 22 km. The smaller and upper of the two barely appears in the heating rate curve, as it falls nto the layer generally characterized by a rapidly decreasing heating rate. While the otal ozone and the maximum concentration were rather less for this ascent than for the earlier one, the maximum heating rate was not very different. Presumably because of the somewhat smaller quantity of ozone above 32 km, the high-level cooling rate was con- siderably less on 2 May than on 14 April.

The radiative divergences we have calculated for the ozone band in the 10-30 km region are about twice as great as those calculated by Plass (1956a) for similar ozone distributions. This difference may partially be because Plass used the Summerfield laboratory data, and we have used the more recent laboratory work of Walshaw. In order to examine to what extent an approximate approach is justified, we carried out a sample calculation using the strong-line approximation and the data of Summerfield (quoted by Plass (1956a), Fig. 1) relating the absorption A by the band to the product pm of pressure and ozone mass. We then used Eq. (12) in the form

$$F = \int_{z'=0}^{z} B(z') \frac{d\tau(z, z')}{dz'} dz'$$

where $\tau = 1 - A$. It is evident that this procedure is extremely simple by comparison to the more exacting one described above. It resulted in heating-rates of about half those calculated by the more exact method, though showing the same general features of positive heating rates in the low stratosphere and negative ones near 30 km. The finer features, for instance the double hump in the divergence curve in the lower stratosphere, were missing from the results of the simple calculation.

5. Discussion

The magnitudes of the heating rates which we have found in the lower stratosphere vary from about + 0·4°C per day to − 0·4°C per day. They are of the same order as those due to carbon dioxide (see Brooks 1958), and thus demonstrate the importance of the radiative transfer in the ozone band. It is not possible, as yet, to make calculations of the total radiation budget in this region of the atmosphere because of the lack of sufficiently accurate data on the contribution of water vapour and of the absorption of solar radiation in this region.

It is also clear from the calculations we have described that any such total radiation study will be hampered by our lack of knowledge of the effective radiative temperature of the troposphere which is so much dependent on cloud distribution and height. The radiative heating of the ozone in the 10-20 km region almost entirely disappears if cloud is present just below the tropopause. As can be seen from Figs. 9 and 12, the error resulting from a wrong effective temperature for the troposphere is probably greater than that resulting through the use of sweeping approximations in the radiation calculations. It is fair to point out, however, that the effect of cloud on the radiation transfer due to carbon dioxide and water vapour will be much less because of the substantial concentrations of these gases in the troposphere; though it may not be negligible.

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Appendix

Computer Time

In the scheme adopted, a matrix consisting of over 8,000 elements and representing *ectroscopic* properties of ozone at some 25 levels was built up in the store of the computer. his calculation required nearly 30-minutes time on the Mercury computer. Weighting ese elements according to the ozone sounding and carrying out the remainder of the lculations required another 30 minutes. Hence one set of heating rates for four base mperatures, required nearly one hour for each spectroscopic sample. For each additional cent, another 30 minutes were required, since the same matrix could be used. Several ternative procedures were considered; some more economical in storage space, which ould have permitted computations on larger samples, but only at a corresponding crease in machine time. Abridged calculation procedures seemed tolerable from the oint of view of precision, but tended to increase traffic time in the machine to produce o important overall saving in time.

Of importance, however, is the possible elimination of the matrix and thus of the ow and poorly accessible store. This can be achieved by calculating the absorption befficient (essentially the sum of some 40 divisions) every time it is needed, with a large esulting saving in computer traffic time. This is feasible if the time for each division can e reduced drastically from the Mercury requirement of 3·6 millisec. A 100-fold reduction promised for the Atlas and other new computers, which might result in an overall 0-fold reduction. In one hour of machine time, one could therefore expect to calculate 0 samples, or with the more relaxed methods of integration (as indicated in the text), 5 spectral samples. It is obvious that a judicious distribution of these through the entire and would permit greatly improved precision in the final result, or – what might be nore important in the end – it would provide an entirely objective standard for judging he spectroscopic models that will still be needed for routine calculations.

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Atmospheric observations of line intensity and half-width in the rotation and ν_2 vibration-rotation bands of water vapour

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(Manuscript received 26 July 1960)

SUMMARY

Curves of growth have been measured in the solar spectrum between 8 and 13 μ on five atmospheric water-vapour lines, three in the rotation band and two in the ν_2 vibration-rotation band. The water vapour in the path, from radiosonde data, was never less than 1 g cm^{-2}, so that it was possible to measure S_0 (the line intensity at $T_0 = 273°K$) and α_0 (the half-width at $p_0 = 1{,}000$ mb) separately only for the weak line at 914·03 cm^{-1}, with errors of approximately 3 and 5 per cent respectively. For four strong lines at 871·32, 878·61, 1106·74 and 1111·53 cm^{-1} only the product $S_0\,\alpha_0$ could be obtained, with an estimated error of 8 per cent. The results of the rotation lines show reasonable agreement with theoretical values. Using theoretical values of α_0, the normalizing factor v for the two ν_2 lines are 0·0165 for 1106·74 and 0·0126 for 1111·53 cm^{-1}, which differ considerably from the overall normalizing factor 0·005 given for this band by Benedict *et al.* (1952) and show a slight dependence on the rotational quantum number J.

1. INTRODUCTION

The intensity S and half-width α of the individual lines of the rotation and ν_2 vibration-rotation bands of water vapour are important parameters in calculating the radiative transfer in the atmosphere. The absolute line intensity S can be calculated from quantum mechanical considerations (Cross, Hainer and King 1944) and is given by

$$S \equiv \int_0^\infty k_\nu \, d\nu = \frac{8\pi^3 N \rho_0^2 (1 - e^{-hc\nu_0/KT}) \nu_0 g_{J_\tau} e^{-E''/KT} |A|^2 v}{3hc\, Q} \quad . \quad . \quad (1)$$

Where

k_ν – monochromatic absorption coefficient

ν_0 – frequency of the line centre (cm^{-1})

N – number of molecules per unit mass

ρ_0 – dipole moment

T – absolute temperature

J_τ – rotational quantum number, associated with pseudo-quantum number τ

g_{J_τ} – statistical weight of the initial state $= (2J + 1) f$, where $f = 1$ or 3 according to whether τ is even or odd

E'' and E' – energies of initial and final states respectively

Q – statistical sum and equal to $\Sigma g_{J_\tau} e^{-(E' - E'')/KT}$

$|A|$ – the matrix element of the directional cosine of the dipole moment associated with the transition from E'' to E'

v – normalizing factor of vibration-rotation transition relative to pure rotation transition

h, K and c – Planck constant, Boltzmann constant and velocity of light, respectively.

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The line strength $(2J + 1)\ |A|^2$ of the water-vapour lines in the rotation band at 50 μ have been computed by Randall *et al.* (1937); Cross *et al.* (1944) and a valuable extension of the latter calculation has been recently published by Schwendeman and Laurie (1958). For the ν_2 band, relative line intensities only can be calculated (Benedict *et al.* 1952). To check the theoretical line intensities of the rotation band and to normalize the lines of the ν_2 band, experimental evidence must be obtained under known environmental conditions.

The half-width and line-shape in the lower atmosphere are mainly determined by collision (pressure) broadening (Goody 1954). The half-width depends on the pressure and collision diameter, the latter depending on the rotational quantum number and so varies from one line to another. Recently Benedict and Kaplan (1959) have calculated the half-width of most lines in the rotation and ν_2 bands of water vapour when broadened by N_2, using Anderson's (1949) theory of collision broadening. Such computations form a landmark in the path towards the solution of radiative transfer problems in the atmosphere, but experimental evidence must also be provided here before confident application is possible.

Very few precise measurements of line intensities have been carried out under atmospheric conditions. Benedict *et al.* (1952) measured, in the laboratory, at relatively high temperature and vapour concentration, the intensities of some lines in the rotation and the ν_2 bands. They inferred a normalizing factor for the ν_2 band by comparision with theoretically calculated intensities. Until recently half-width measurements of individual lines are few owing to the inadequate resolution available. Adel (1947) and Cloud (1952) measured the approximate width of a few lines in the region 16-20 μ but the only accurate measurements ($\pm$ 10 per cent) obtained, so far, are on the micro-wave line (6_{-5}, 5_{-1}, at 0·744 cm^{-1}) by Becker and Autler (1946).

This paper describes the measurements of S_0 (the line intensity at $T_0 = 273°K$) and α_0 (the half-width at $P_0 = 1{,}000$ mb) or their product, in the solar spectrum of three lines in the rotation and two lines in the ν_2 band, using a grating spectrometer (resolution 0·3 cm^{-1} at 1,000 cm^{-1}). The measurements, which were carried out at the Imperial College Field Station at Ascot, will be discussed and compared with theoretical computed values from earlier papers referred to.

2. Theory of the method used

The equivalent width W of a single line is given by

$$W = \int_{-\infty}^{\infty} (1 - e^{-k_\nu M})\, d(\nu - \nu_0) \quad . \quad . \quad . \quad (2)$$

where M is the amount of absorbing material in the path of radiation.

The monochromatic absorption coefficient k_ν for a line will be assumed to be given by Lorentz formula

$$k_\nu = \frac{1}{\pi} \frac{S\alpha}{(\nu - \nu_0)^2 + \alpha^2} \quad . \quad . \quad . \quad . \quad (3)$$

In the solar spectrum, however, $S = S(T)$ and $\alpha = \alpha(P, T)$ and must be taken into account.

The temperature dependence of S is given from Eq. (1) ignoring the term $(1 - e^{-hc\nu_0/KT})$, which is nearly unity, by

$$S(T) \propto \frac{e^{-E''/KT}}{Q} \quad . \quad . \quad . \quad . \quad . \quad (4)$$

where the statistical sum Q is proportional to $1/T^{3/2}$ for asymmetric top molecules (Herzberg 1951). The temperature effect is introduced into M to produce M^* the scaled water vapour in the path and the line intensity S_0 at 273°K is taken as a constant.

The half-width of a line in the troposphere for collision broadening follows

$$\alpha \propto PT^{-n} \quad . \quad . \quad . \quad . \quad . \quad (5)$$

where n differs from one line to another, and for the lines chosen, is less than 0·4 (Benedict and Kaplan 1959). However, the temperature in the troposphere changes little ($\pm$ 10 per cent) compared with change in pressure, so that its effect on α can be ignored. The existence of pressure gradient along the path introduces a complication which has only been solved for special cases (Plass 1956). However, Curtis (1952) has shown that, in general, a good approximation is obtained by considering all the absorption to take place at a mean half-width $\bar{\alpha}$ given by the mean pressure $\bar{P}$

where
$$\bar{P} = \frac{\int P\, dm^*}{\int dm^*} \quad . \quad . \quad . \quad . \quad . \quad (6)$$

and m^* = scaled water vapour in the zenith.

For the computation of m^* and $\bar{P}$ the atmosphere is divided into nine layers with boundaries at 1,000, 950, 900, 850, 800, 750, 650, 550, 450 and 350 mb. A small correction was afterwards applied for the variation of the surface pressure from 1,000 mb. Above 350 mb the amount of water vapour is less than 0·6 per cent of the total amount and its contribution to the absorption, bearing in mind the temperature effect is less than 0·1 per cent.

The integration of Eq. (2) with Eq. (3) and the introduction of the temperature and pressure effect gives the well-known Ladenburg and Reiche (1913) equation

$$W = 2\pi\, \alpha_0 \frac{\bar{P}}{P_0} X e^{-X} [I_0(iX) - iI_1(iX)] \quad . \quad . \quad . \quad (7)$$

where $X = \dfrac{S_0 M^*}{2\pi\alpha_0 (\bar{P}/P_0)}$, I_0 and I_1 are Bessel functions of the order 0 and 1.

(*i*) for small X $\qquad W = S_0 M^* \quad . \quad . \quad . \quad . \quad . \quad (8)$

(*ii*) for large X $\qquad W = 2\sqrt{S_0 \alpha_0 (\bar{P}/P_0) M^*} \quad . \quad . \quad . \quad (9)$

Case (*i*) is known as the linear, and case (*ii*) as the square-root region of absorption. A plot of $\log (W/M^*)$ against $\log (M^* (P_0/\bar{P}))$ gives the absorption 'curve-of-growth' with slope approaching zero in the linear region (8) and $-\frac{1}{2}$ in the square-root region (9). Re-writing Eq. (7) as

$$Y = \frac{W}{2\pi\alpha_0 (\bar{P}/P_0) X} = \frac{W}{S_0 M^*}$$

$$Y = e^{-X} [I_0(iX) - iI_1(iX)] \quad . \quad . \quad . \quad (10)$$

so that the curve of $\log Y$ against $\log X$ is a universal curve. Kaplan and Eggers (1956) have published a table of the function $X e^{-X} [I_0(iX) - iI_1(iX)]$ from which Y can easily be calculated. The curve of Eq. (10) was plotted on tracing paper and the best fit for an observed set of values of (W/M^*) and $(M^* (P_0/\bar{P}))$ was obtained by adjusting the curve to the observed points, thereby obtaining the two parameters S_0 and S_0/α_0. Their variances were found by fixing S_0 and varying S_0/α_0 and vice versa, until the least variance was obtained.

Note that when $\log Y = 0$

$$\log S_0 = \log \frac{W}{M^*} \quad . \quad . \quad . \quad . \quad . \quad (11)$$

and when $\log X = 0$

$$\log \alpha_0 = \frac{S_0 M^*}{2\pi} \frac{P_0}{\bar{p}} \quad . \quad . \quad . \quad . \quad (12)$$

When the observations were restricted to the square-root region, as was the case for four of the lines used, W was plotted directly against $(M^* (\bar{P}/P_0))^{\frac{1}{2}}$ and the best line through the origin was found by the least-square method. The slope of the line gives $2\sqrt{S_0 \alpha_0}$.

(*a*) *Absorption by wings*

In practice, the line area (Eq. 2) could only be measured to relatively small frequency range from the centre of the line. The contribution to the equivalent width from wing absorption is given by

$$W' = 2\int_{\delta\nu}^{\infty} (1 - e^{-k_\nu M})\, d(\nu - \nu_0) \quad . \quad . \quad . \quad (13)$$

where $2\,\delta\nu$ is the frequency interval over which the line area has been measured. For a Lorentz line shape it can be shown (Benedict *et al.* 1956) by direct integration of (13) with (3) that a first-order approximation

$$W' = W\frac{2\alpha_0}{\pi\,\delta\nu} \quad \text{for linear region} \quad . \quad . \quad . \quad (14)$$

$$W' = \frac{W^2}{2\pi\,\delta\nu} \quad \text{for square-root region} \quad . \quad . \quad . \quad (15)$$

An iteration method, which converges very quickly, is used to calculate the correction to the measured equivalent width. This method of obtaining the wing contribution is an approximation because the calculation assumes that the background has been drawn between two points so far away from the centre of the considered line that the absorption due to it is negligible; and the line has a sharp cut-off. In fact both assumptions are wrong and in such direction as to make the actual correction larger than estimated by 1-2 per cent.

3. Choice of lines and observational procedure

Ideally each line selected for this experiment should be:

(*i*) remote from any other line

(*ii*) a single line and not a blend

(*iii*) of medium intensity (the growth curve would then extend from the linear to the square-root region, so that S_0 and α_0 could be obtained separately).

These conditions are very restrictive and the best obtainable lines are listed in Table 1 and shown in Fig. 1.

TABLE 1. Details of lines chosen

Line number	cm^{-1}	Note
1	871·32	Single line, fairly strong
2	878·61	Blend with weaker component of H_2O
3	914·03	Single rather weak line
4	918·53	Blend with CO_2 component
5	1,106·74	Single fairly strong line
6	1,111·53	Single fairly strong line

Line 2 is a blend of two water-vapour lines, but the weaker component is only 6·1 per cent of the intensity of the stronger component (Benedict *et al.* 1952). The results given later in Table 2 for this line, refer to the stronger component, a correction having been made for the weak line. Line 4 contains a CO_2 component at 918·76 cm^{-1} and since no data were available on this line it proved useless in the ultimate analysis.

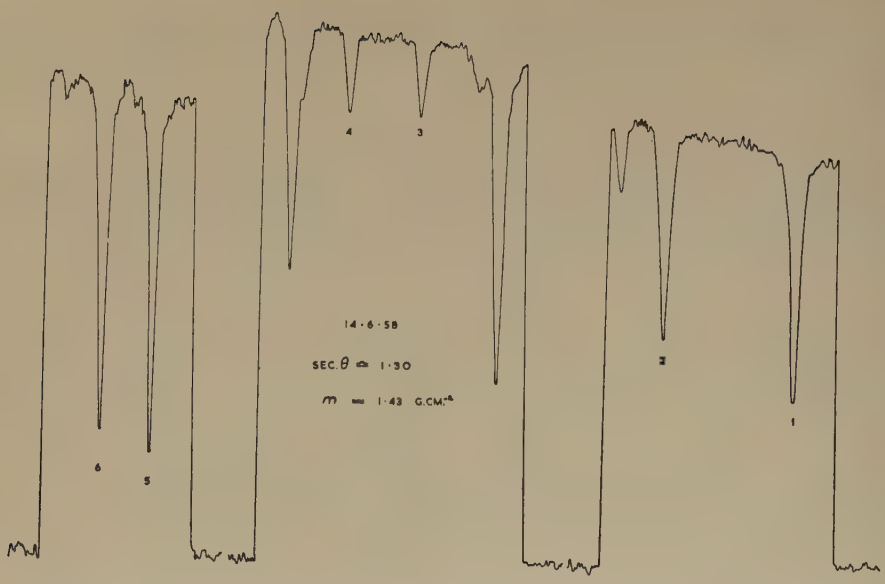

Figure 1. A typical record of solar absorption showing all lines, zero, and noise level.

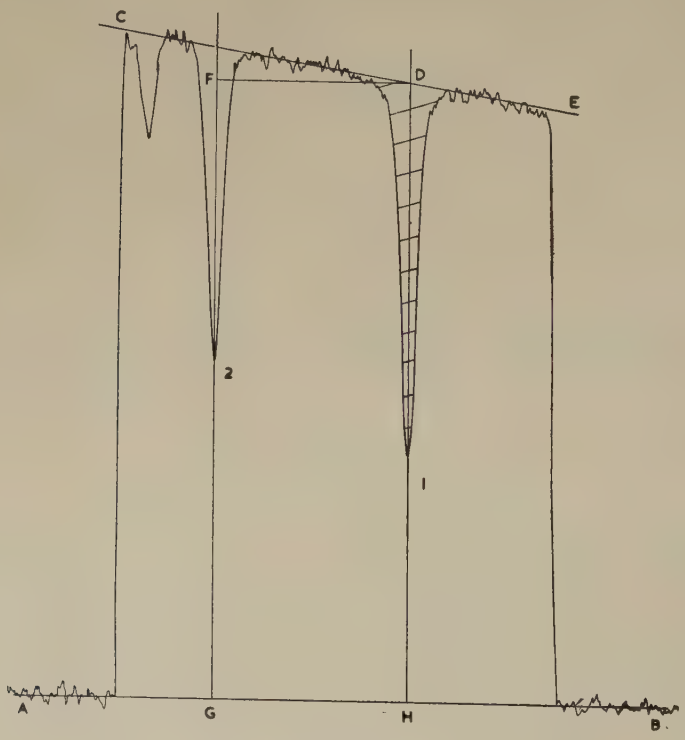

Figure 2. Method of evaluating the equivalent width.

AB = zero line CDE = background

DH, FG are drawn perpendicular to AB through line centres

DF parallel to AB

$$\text{Equivalent width } W = \frac{\text{shaded area}}{\text{area DFGH}} \times \text{frequency corresponding to GH.}$$

The spectrometer and recorder and the linearity of the system, better than 0·3 per cent over the range of signal used, have been described by Saiedy and Goody (1959). The sun was used as the source of radiation and the airmass value, corrected for earth curvature and air refraction, was calculated from the time of observation. Although the value obtained for the equivalent width W is independent of the slit-width, it is found, in practice, preferable to use the narrowest slit compatible with good signal to noise ratio, especially when the absorption is small. A slit-width of approximately 0·4 cm^{-1} was found to be suitable. The scanning speed was limited by the response time of the detector system and was such that a line was traversed in approximately 15 sec. The observations were taken between February 1958 and the summer of 1959, with different airmass values, thus giving a large range of water vapour in the path, i.e., 1 to 30 g cm^{-2}. The evaluation of the equivalent width W from the records is made clear in Fig. 2, a planimeter being used to obtain the areas out to approximately $\pm$ 1·5 cm^{-1} from the line centre, and correction (10 per cent) for the absorption in the wings of the line was applied according to Section 2 (*a*).

4. Results

The growth curves for the lines are shown in Fig. 3. The results computed from these curves are shown in Table 2 for both rotation and ν_2 lines. Theoretical values of S_0 were calculated using Schwendeman and Laurie's (1958) tables. The statistical sum and dipole moment used here are those of Benedict *et al.* (1952) and are 178 at 287°K and 1·86 $\times$ 10^{-18} e.s.u. respectively. The theoretical values of α_0 are those of Benedict and Kaplan (1959).

TABLE 2. Observed and theoretical line parameters

						Measured			Theoretical		
Line	ν_0 cm^{-1}	Transition J_τ'	J_τ''	Band	Sub-branch	S_0 $cm^{-1}/(g\ cm^{-2})$	α_0 cm^{-1}	$S_0\alpha_0$ g^{-1}	S_0 $cm^{-1}/(g\ cm^{-2})$	α_0 cm^{-1}	$S_0\alpha_0$ g^{-1}
1	871·32	11_{-2}	10_{-10}	Rot.	$R_{5,-3}$			0·0231 $\pm$ 16	0·444	0·0469	0·0208
2	878·61	12_{-4}	11_{-10}	Rot.	$R_{3,-3}$			0·0127 $\pm$ 7	0·177	0·0674	0·0119
3	914·03	13_5	12_{-1}	Rot.	$R_{3,-3}$	0·031 $\pm$ 1	0·047 $\pm$ 2	0·00147 $\pm$ 7	0·036	0·0445	0·00160
5	1106·74	19_7	10_{-3}	ν_2	$P_{-3,1}$			0·0583 $\pm$ 56		0·0525	
6	1111·53	6_5	7_1	Rot.	$P_{-3,3}$			0·0617 $\pm$ 62		0·0757	

Before discussing the results we enquire whether the scatter of points about the mean curves can reasonably be accounted for by experimental error. The error of an individual measurement depends on (a) error in measuring the water-vapour content in the path from radiosonde data, (b) error in planimetering the line, i.e., measuring the equivalent width W and (c) instrumental errors. An objective method was used to estimate the error due to each source, and the final results were compared with the standard deviation calculated from the curve of growth.

(a) The water-vapour content was deduced from observations of three radiosonde stations surrounding Ascot (Crawley, Hemsby and Camborne) on eleven days when the meteorological conditions for all stations were essentially the same. The standard deviation in the measurement of water vapour was found to be $\pm$ 20 per cent. There is, undoubtedly, a small local component in the variation of the water vapour which is very difficult to assess. An error of 15 per cent in the measurements of water-vapour content from radiosonde ascents is not an unreasonable estimate.

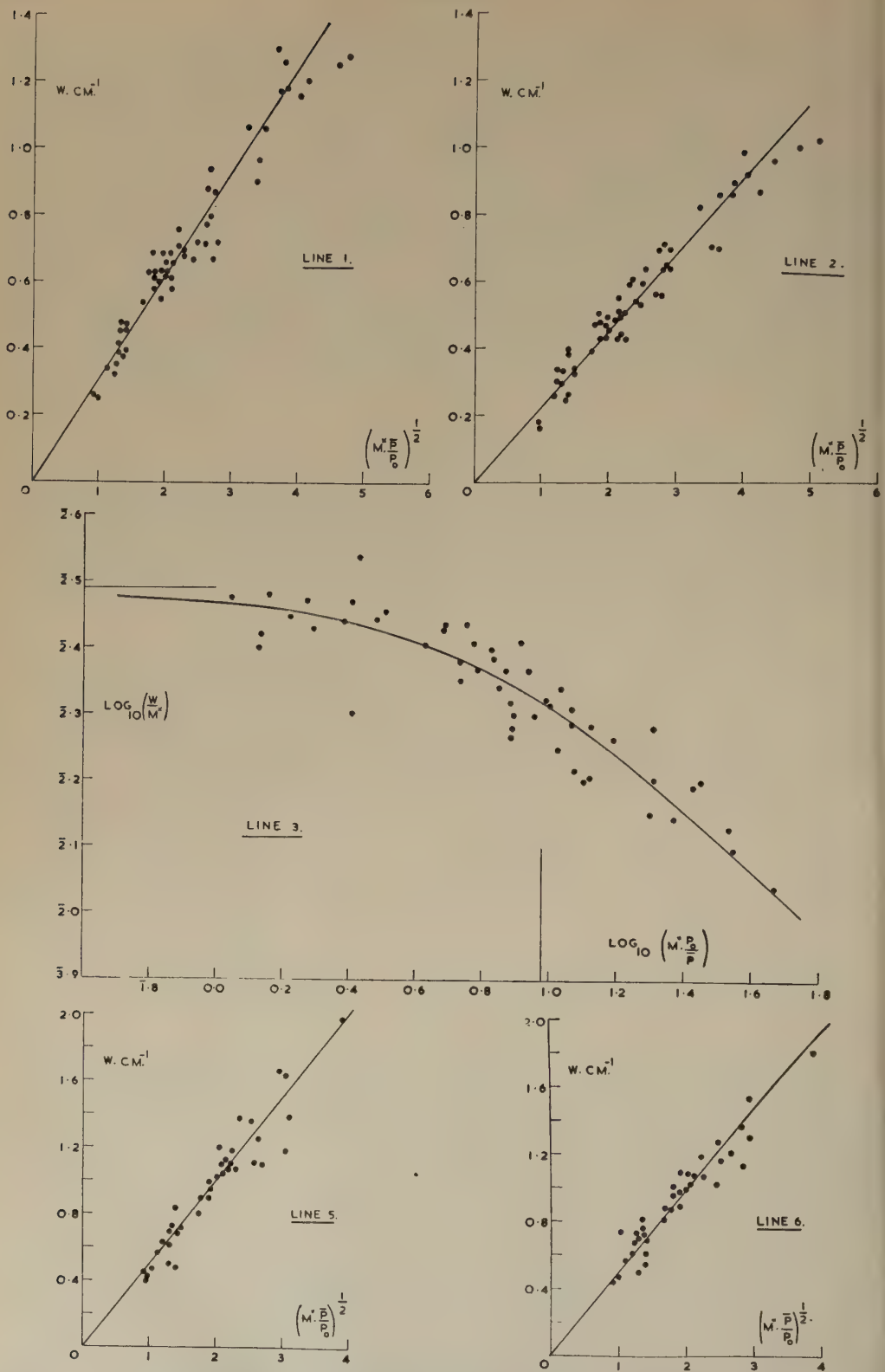

Figure 3 The curves of growth of the five lines analysed. (Note that only line 3 is sufficiently weak to enter the linear region).

(b) The assessment of error in measuring the equivalent width W depends on setting the background line (i.e. the level of zero absorption) and also on the size of the area to be measured. When the amount of water vapour is large (say, 20 g cm^{-2}) the area is large and the error in measuring it was not more than 2 per cent, but the actual setting of the background is much more uncertain owing to absorption by neighbouring lines. However, at small water-vapour content (2 g cm^{-2}) the reverse is true; the background can be set with certainty but because the area is so small, particularly for weak lines, measurement is very difficult. The smallest area measured was for line 3 (0·04 cm^{-1}) with an error of $\pm$ 15 per cent. A test of the reproducibility of the equivalent width was carried out on lines 1 or 2 for medium absorption ($W \simeq 0{\cdot}2$ cm^{-1}). These lines were scanned near noon when the airmass is hardly changing. The standard error found was approximately $\pm$ 4 per cent.

(c) Instrumental errors could be introduced if the driving speed of the grating is not constant (the recording-paper speed was found to be constant to better than $\pm$ 0·3 per cent). This was checked by introducing calibrating pips at small intervals, approximately 1 cm^{-1} apart, on the driving gear, and it was found that the error introduced from variable-drive speed did not amount to more than 1 per cent and may therefore be ignored.

TABLE 3. OBSERVED ERROR COMPARED WITH INDEPENDENTLY ESTIMATED ERROR FROM VARIOUS SOURCES

Line	b	σ_b	σ_γ	$(\sigma_\gamma)'$
1	0·2845	0·0102	0·071	0·053
2	0·2068	0·0058	0·046	0·043
5	0·4781	0·0238	0·103	0·085
6	0·4445	0·0237	0·095	0·074

b = slope of regression line

σ_b = standard error of b, hence in $\sqrt{S_0\,\alpha_0}$

σ_γ = standard deviation of an individual value of W about the regression line in cm^{-1} from regression analysis

$(\sigma_\gamma)'$ = as for σ_γ but given by the combination of the independently estimated errors in $(M^*)^{\frac{1}{2}}$ and W

Summarizing, the independent estimates of the main source of errors are 15 per cent and 4 per cent. We now compare these estimates with observed errors from Fig. 3. For lines in the square-root region, straightforward linear regression analysis can be applied. The detailed results of this analysis are shown in Table 3 from which it is seen that the independent estimates of errors account adequately for all observed scatter. For line 3, which extends into the linear region and for which the parameters are fitted by the method described in Section 2, the estimated standard deviation of the ordinate $\log_{10}(W/M^*)$ of the individual points was found to be 11 per cent and it can be shown that this error can be accounted for, as for the other lines, from the experimental error mentioned above. However, for the estimation of the error in S_0 and α_0 separately, no straightforward analysis is available and the following procedure was adopted. The accuracy with which S_0 and α_0 can be determined from an individual observed point depends on the parameter X, where $X = \dfrac{S_0\,M^*}{2\pi\alpha_0\,P/P_0}$. For small X ($X < 0{\cdot}5$), W/M^* is a function of S_0 (Eq. 8), while for large X ($X > 2$), it is a function of $\sqrt{S_0\,\alpha_0}$ (Eq. 9). To determine the accuracy of S_0 and $\sqrt{S_0\,\alpha_0}$ the observations were divided arbitrarily into three groups according to the values of X, and the following weights were assigned to each :

	Weights	
X	S_0	$\sqrt{S_0\,\alpha_0}$
0 -0·5	1	0
0·5-2·0	$\frac{1}{2}$	$\frac{1}{2}$
$> 2{\cdot}0$	0	1

With these weights the standard deviation of S_0 for line 3 was found to be $\pm$ 3 per cent and $\sqrt{S_0\,\alpha_0}$ to be 2·3 per cent, giving an error of approximately 5 per cent in $S_0\,\alpha_0$.

5. Discussion

The values of S_0 and α_0 for line 3 show agreement within 14 per cent and 15 per cent respectively, with the theoretical values (Table 2). Note especially that the theoretical value of the half-width agrees with the experimental value within the uncertainty of the latter. This is regarded as providing strong support for the theoretical derivation by Benedict and Kaplan (1959) of the half-width due to H_2O-N_2 broadening and its applicability to the atmosphere with its mixture of constituents of N_2, O_2 and H_2O. It has been argued by Benedict and Kaplan (1959) that the half-width of H_2O broadened by air is less than the half-width broadened by N_2. However, the halfwidth of H_2O broadened by H_2O is much greater than either (5 times larger; see Becker and Autler (1946)) and although the content of water vapour in the atmosphere is small, it could, theoretically, off-set the reduction due to the presence of O_2.

With this notable agreement between theory and observation, we are encouraged to take the calculated half-width to be substantially correct and infer line intensities for four lines in the square-root region from the observed values of $S_0\,\alpha_0$. These values are shown in Table 4 together with the theoretical values given in Table 2 and discussed in Section 4.

TABLE 4. Line intensities obtained from measurements of $S_0\,\alpha_0$ compared with calculated rotational line intensities

Line	Band	Inferred S_0 $cm^{-1}/(g\ cm^{-2})$	Theoretical S_0 $cm^{-1}/(g\ cm^{-2})$
1	Rot.	0·491 ± 0·034	0·444
2	Rot.	0·189 ± 0·011	0·177
5	ν_2	1·11 ± 0·11	
6	ν_2	0·816 ± 0·082	

The ν_2 vibration-rotation line intensities, calculated on a pure rotation basis, and the experimental values are given in Table 5 together with the normalizing factor v (Eq. 1) deduced from the intensities.

TABLE 5. The normalizing factor v for the two lines measured in the ν_2 band

Line	Band	Inferred S_0 $cm^{-1}/(g\ cm^{-2})$	Theoretical S_0 on pure rot. assumption	v	J''
5	ν_2	1·11	67·3	0·0165	10_{-3}
6	ν_2	0·816	64·7	0·0126	$7_{\ 1}$

Benedict *et al.* (1952) gave an overall normalizing factor equal to 0·005 for the whole ν_2 band but added that it was too small by factor of 3 in the region 900-1,050 cm^{-1}. Our inferred values of v confirm this statement and they may also be taken to show some dependence of v on the rotational quantum number J (see Table 5). Observations on additional lines of different quantum numbers are necessary in order to determine the behaviour of v and to assess its physical significance. It is also desirable to extend these measurements into the linear region so that the line intensities and half-width can be determined separately. A horizontal-path experiment is now being undertaken to supplement the solar observations reported here, and to provide the additional data.

In conclusion, the theoretical values of S_0 for the rotation band as deduced from Schwendeman and Laurie's (1958) tables of line strengths and the half-width α_0 as given by Benedict and Kaplan (1959) for the rotation band of water vapour, are in good agreement with observed values of S_0 and α_0 in the atmosphere. The position regarding the ν_2 line intensities appears to be less satisfactory; relative values calculated on a rotational basis can only be regarded as rough approximations. The variation of v with J must be established before actual line intensities in this band can be calculated.

Acknowledgments

I am deeply indepted to Professor P. A. Sheppard and Professor R. M. Goody for their helpful advice. My thanks also go to Dr. C. D. Walshaw and Mr. K. Bignell for many helpful discussions, and to the technical staff, Mr. D. A. Rogers and Mr. A. Pusey, at the Imperial College Meteorological Field Station.

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Vertical series of sea and air temperatures close to the surface, taken aboard the yacht *Petula* in the tropical Atlantic

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(Manuscript received 17 January 1961; in revised form 17 April 1961)

Summary

Sea and air temperatures were taken at a number of points within 2 m of the surface. In general, both sea and air temperature profiles were almost isothermal at more than 1 m from the surface. Air temperatures gave some indication of a change from neutral to stable conditions above about 55 cm. There was little confirmation of an earlier suggested similarity between profiles of wind, temperature and humidity under near-neutral conditions.

1. Introduction

Investigations of the thermal exchanges between the ocean and the atmosphere are invariably handicapped by insufficiency of data; and while recent interest has stimulated the collection of observations, large areas of the globe remain where the essential records of sea and air temperatures close to the surface are lacking.

An important difficulty in the collection of such observations on the high ocean arises from the proximity of recording thermometers to the ship. It has more than once been shown that shipboard heat losses interfere with the accuracy of near-surface thermometer readings to the extent that valid calculations of air-sea energy exchanges cannot be made from them without resorting to doubtful corrections. Charnock (1951) has reviewed various methods devised to overcome this difficulty; more recently Shellard (1958) has described a thermobuoy employed by O.W.S. *Weather Recorder* which secured results in the Atlantic at 50°N before technical faults interrupted a promising programme; and reports by Takahashi (1959) and others, to the First International Oceanographic Congress indicate a widening attack on the problem.

2. Experimental procedure

The transatlantic voyage of the yacht *Petula* in the winter of 1953-4 afforded unusually favourable opportunity for the measurement of near-surface temperatures. The *Petula*, a 12-ton yawl, drifted from Dakar, W. Africa, to Barbados, W. Indies, in a voyage occupying 12 weeks. Along her route oceanographical and meteorological observations were systematically conducted (Evans 1955; Dickson and Evans 1956). A small raft was towed astern, and it was from here that sea and air temperature measurements were made. The raft was built of wood, having as its basis three Admiralty Carley floats. When completed it was 16 ft long including a false bow, 8 ft wide, with a draught and freeboard each of about 6 in. It was towed on 90 ft of the ship's anchor cable. Since the Atlantic crossing was made before the wind, the raft lay continually to windward of the ship.

Aboard the raft, sea temperatures were taken by means of Symons earth thermometers (lagged-bulb thermometers). Five of these were clamped to a wooden pole. Through a hole in the raft's deck the pole was lowered vertically into the sea until a mark near one end just touched the surface. The five thermometers were then at depths of 10 cm, 25 cm, 50 cm, 1 m and 2 m respectively. The pole was held in position for 30 min, drawn up and the temperatures read off.

Air temperatures, recorded immediately after the sea temperatures, were taken with a hand-driven Assmann psychrometer operated at 10 cm, 25 cm, 50 cm, 1 m and 2 m above the sea surface. Trials at various points around the raft showed that the raft's presence had no noticeable effect on the temperatures observed.

All the thermometers used were calibrated in °F and read to the nearest 0·1°. The time-constant of the Symons earth thermometers was not known, but from the description given in *M.O. 577* (1956) it might be assumed to be about 5 min; thus a 30-min submersion would be adequate to indicate temperatures which, in general, varied little from hour to hour. Under the conditions found it is believed that temperatures were correct to an accuracy of $\pm$ 0·2°F. Air temperatures were better, with a believed accuracy of $\pm$ 0·1°F. Great care was taken to ensure that the wet-bulb was free from salt contamination. It should be noted that the figures in Table 1 do not imply an accuracy of 0·01°A.

Observations were made at 40 stations across the Atlantic between longitudes 24°W and 51°W, within the latitudes of 12½°N and 15½°N. They were made at about 1500 hr local time, on days when sea conditions permitted the observer (C. N. D.) to make a passage from the yacht to the raft, that is, in general, in winds up to force 6.

3. Discussion of results

Results have been assembled in 8 groups (Table 1); each of the first six containing five observations, the seventh group, three, and the eighth group, seven. In this way observations in each group occur on successive days, or nearly so. The eighth group gave results differing in form from the seven earlier ones (see below); for these groups the air temperature profile above 1 m is almost isothermal (Fig. 1) and similarly, the sea temperature profile below 1 m.

TABLE 1. Means of sea, air and wet-bulb temperatures (t_s, t_a, t_w) in °A

Group	Date 1953-4	Mean position	Mean wind speed (kt)	Mean wave height (ft)		Distance from sea surface (cm) 10	25	50	100	200
1	Dec. 10-15	13°N 25°W	13	8	t_s	298·62	298·61	298·74	298·82	298·66
					t_a	297·87	297·70	297·60	297·55	297·58
					t_w	295·49	295·16	294·96	294·78	294·71
2	Dec. 16-20	14°N 28°W	6	10	t_s	298·44	298·62	299·18	299·00	298·86
					t_a	297·63	297·40	297·29	297·17	297·13
					t_w	294·91	294·42	293·98	293·64	293·63
3	Dec. 21-25	14°N 30°W	13	8	t_s	298·51	298·60	298·62	298·63	298·58
					t_a	297·57	297·40	297·30	297·20	297·19
					t_w	294·93	294·73	294·60	294·30	294·13
4	Dec. 26-30	15°N 33°W	9	5	t_s	298·14	298·13	298·04	298·04	297·93
					t_a	297·64	297·36	297·27	297·19	297·08
					t_w	294·78	294·40	294·14	293·91	293·79
5	31 Dec. to 4 Jan.	15°N 35°W	9	5	t_s	298·29	298·36	298·43	298·41	298·53
					t_a	297·77	297·66	297·50	297·48	297·50
					t_w	295·36	295·09	294·81	294·58	294·44
6	Jan. 5-9	15°N 37°W	14	10	t_s	298·30	298·31	298·37	298·38	298·37
					t_a	298·16	297·98	297·96	297·84	297·73
					t_w	296·24	295·92	295·67	295·48	295·33
7	Jan. 14-16	14°N 40°W	10	7	t_s	298·24	298·31	298·37	298·22	298·20
					t_a	297·77	297·59	297·46	297·42	297·48
					t_w	295·24	294·88	294·49	294·27	293·98
8	28 Jan. to 3 Feb.	13°N 49°W	11	5	t_s	298·89	299·02	298·99	298·98	298.98
					t_a	298·57	298·40	298·39	298·39	298·42
					t_w	296·43	296·24	296·06	295·67	295·63

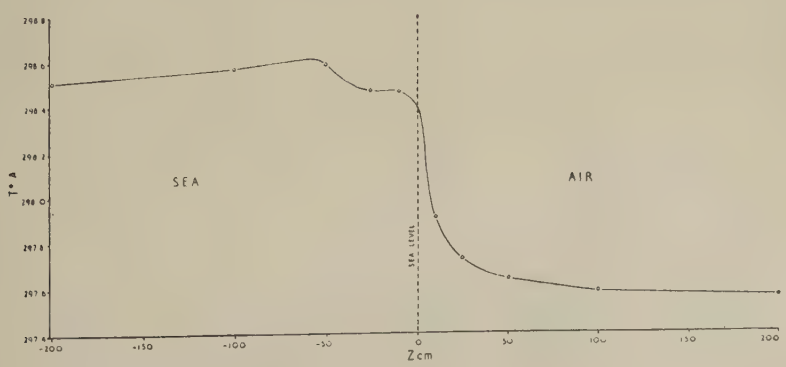

Figure 1. Profile of mean temperatures across the sea surface, derived from the first seven groups of observations.

Fig. 2 shows $t_0 - t_a$ (air temperature differences) against $\log_{10} Z$ (height). Fig. 3 shows $e_0 - e_a$ (saturated vapour pressure differences) against $\log_{10} Z$. Here $e_0 = 0{\cdot}98 \times$ S.V.P. at the sea-surface temperature, 0·98 being a salinity correction. The straight portions of all these curves were fitted by the method of least squares. Equations for these straight portions are given in Table 2 together with equations for $t_0 - t_w$ (wet-bulb differences; not figured). Values for t_0 and e_0 were obtained by extrapolation.

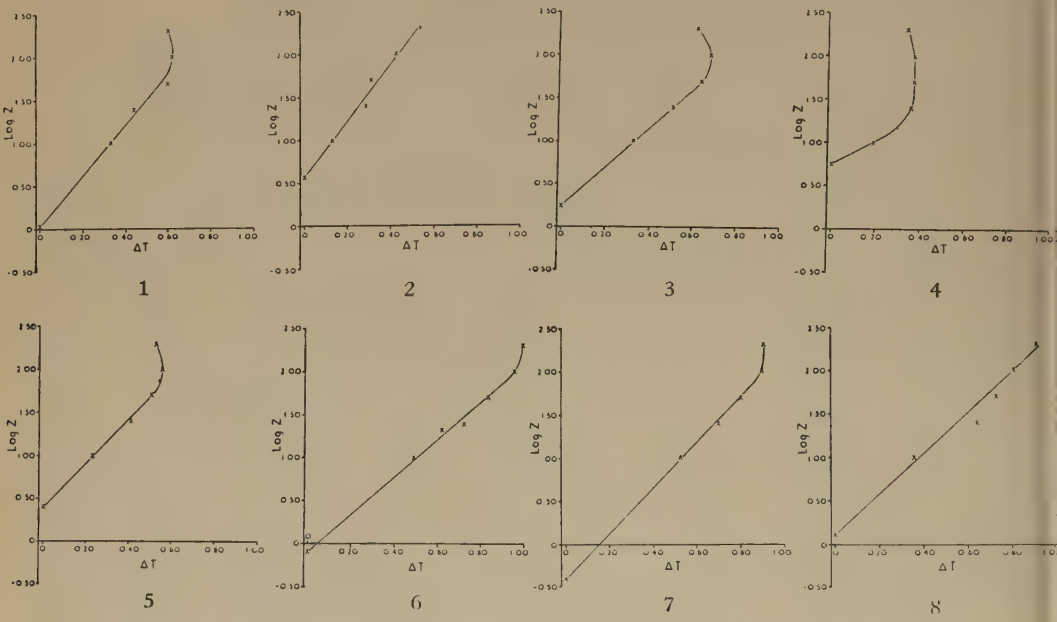

Figure 2. $\Delta T = t_0 - t_a$, plotted against $\log_{10} Z$.

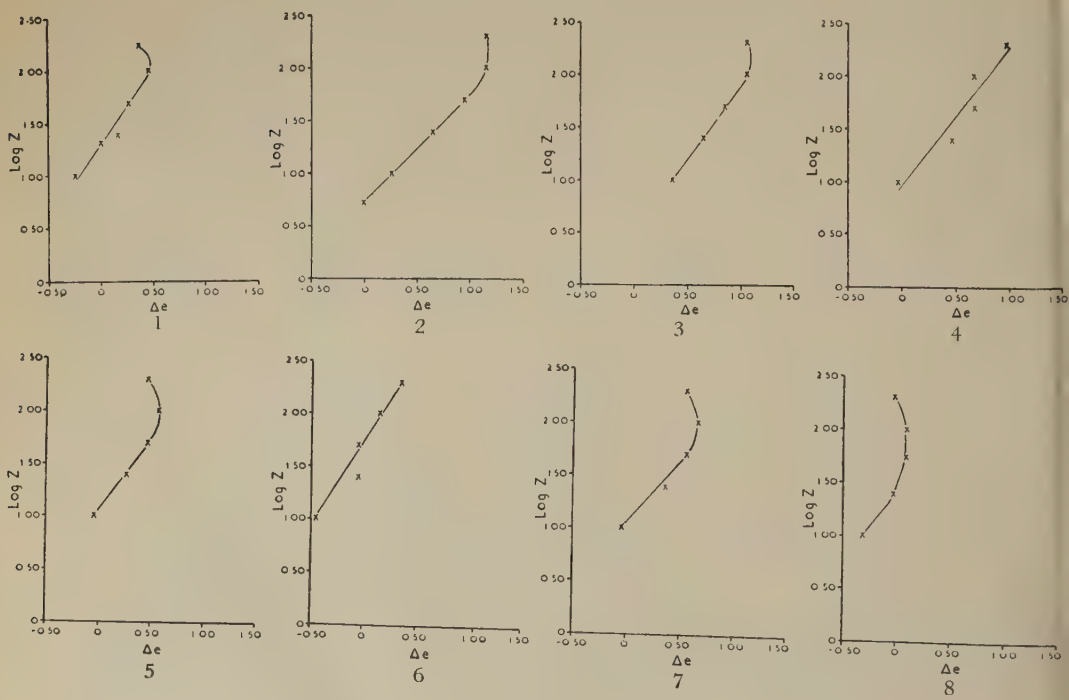

Figure 3. $\Delta e = e_0 - e_a$, plotted against $\log_{10} Z$.

TABLE 2. VALUES FOR $t_0 - t_a$, $e_0 - e_a$ AND $t_0 - t_w$ IN TERMS OF $\log_{10} Z$

Group 1	$0{\cdot}400\ (\log Z - 0{\cdot}413)$	$0{\cdot}707\ (\log Z - 1{\cdot}330)$	$0{\cdot}495\ (\log Z + 2{\cdot}85)$
2	$0{\cdot}460\ (\log Z + 0{\cdot}100)$	$0{\cdot}975\ (\log Z - 0{\cdot}725)$	$1{\cdot}263\ (\log Z + 1{\cdot}55)$
3	$0{\cdot}379\ (\log Z + 0{\cdot}400)$	$0{\cdot}688\ (\log Z - 0{\cdot}525)$	$0{\cdot}654\ (\log Z + 3{\cdot}78)$
4	$0{\cdot}426\ (\log Z - 0{\cdot}100)$	$0{\cdot}867\ (\log Z - 0{\cdot}963)$	$0{\cdot}786\ (\log Z + 2{\cdot}50)$
5	$0{\cdot}340\ (\log Z - 0{\cdot}038)$	$0{\cdot}714\ (\log Z - 1{\cdot}050)$	$0{\cdot}740\ (\log Z + 2{\cdot}70)$
6	$0{\cdot}304\ (\log Z - 0{\cdot}563)$	$0{\cdot}625\ (\log Z - 1{\cdot}730)$	$0{\cdot}725\ (\log Z + 1{\cdot}80)$
7	$0{\cdot}446\ (\log Z - 0{\cdot}250)$	$0{\cdot}888\ (\log Z - 1{\cdot}630)$	$0{\cdot}965\ (\log Z + 1{\cdot}96)$

The results give some indication of a deviation from neutral stability at about 55 cm; above this height stable conditions prevail. By contrast, work reviewed by Charnock (1951), e.g., aboard the *Altair*, has suggested that the region of neutral stability sometimes extends to greater heights than this, and individual results aboard the *Petula* support this view. However, it should in fairness be stated that the results presented here must be interpreted with some reserve, in view of the conditions under which they were obtained.

Shellard (1958) examined the wind profile immediately over an aerodynamically rough surface under neutral conditions, for which he cites the equation $U/U_* = k^{-1} \log (Z/Z_0)$ where U_* is the friction velocity (defined by $U_* = \sqrt{\tau/\rho}$ where τ_0 is the surface drag and ρ, the air density), k is von Karman's constant and Z_0 is the roughness parameter which characterises the geometry of the boundary.

From weather ship observations Shellard obtained values of z_0 of about $10^{-2{\cdot}8}$ and $10^{-3{\cdot}4}$ cm. He suggests a possible relationship between this quantity and another, Z_0, which he defines as the intercept of the temperature or humidity profile on the log Z axis. On 9 July 1956, when z_0 was $10^{-3{\cdot}4}$ cm, Z_0 was found to be $10^{-2{\cdot}1}$ cm for temperature and $10^{-2{\cdot}7}$ cm for vapour pressure.

The *Petula* observations are more numerous and varied. In group 2 and 3, Z_a (intercept of the air temperature curve on the log Z axis) have values of 10^{-1} and 10^{-4} respectively, corresponding approximately with Shellard's figures. The remaining groups are, however, considerably greater, lying between $10^{0{\cdot}1}$ and $10^{0{\cdot}563}$. The constant term outside the brackets (see Table 2) is about twice Shellard's value.

Z_e (intercept of the vapour pressure curve on the log Z axis) has a mean value of 10^1; the spread of the constant terms is wider than for Z_a. For Z_w (intercept of wet-bulb temperature curve on the log Z axis) the figures are more constant. Only here does the mean value of $10^{-2{\cdot}2}$ approximate to Shellard's value of $10^{-2{\cdot}7}$. Thus, the similarity between Zo, Za and Ze appears somewhat questionable.

The eighth group of observations aboard the *Petula* is noteworthy in that there is no indication of neutral conditions. It is possible that the proximity of the American land-mass had a stabilising effect on the air, but an adequate consideration of the problem must await further observations.

ACKNOWLEDGMENT

The authors are indebted to M. J. Blackwell who kindly read the manuscript.

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Point discharge from natural and artificial points (Pt. II)

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(Manuscript received 5 January 1961)

Summary

Measurements have been made of the point-discharge current through an artificial point at about tree-top height, the potential gradient at the ground nearby and the wind speed. The results are similar in form to those for an isolated point at greater height and do not show the square-law relation between current and potential gradient that has previously been found when wind-speed has not been taken into account. The current flowing down a living tree has also been measured and obeys a similar relation with potential gradient.

1. Introduction

There have been a number of investigations of the relation between the point-discharge current through a point erected in the atmosphere and the potential gradient at the ground nearby. These can be divided into two categories : (1) investigations with points at about tree-top heights, giving point discharge only in stormy weather, and for which the wind-speed has not been measured; and (2) investigations with isolated points, at heights well above the tree tops, giving point discharge not only in stormy weather but often also in non-stormy weather, wherein the wind-speed is an important factor.

The results from the two types of investigation have been expressed in formulae which differ considerably, one containing the wind-speed and the other not. It therefore seemed worth while to make fresh measurements with a point at tree-top height but including measurements of wind-speed, in order to ascertain whether the same formula will fit these results as is adequate for isolated points.

The investigations so far discussed have been carried out with artificial metal points, but no similar investigation has been attempted for the point-discharge current down natural trees, except that of Schonland (1928) with a tree cut down and supported on insulators and that of Maund and Chalmers (1960) by an indirect method. It seemed worth while to attempt to make measurements of this kind with a living tree.

2. Previous results

The results for points at about tree-top height have been expressed in the form :

$$I = A\,(F^2 - M^2)^* \quad . \quad . \quad . \quad . \quad . \quad (1)$$

where I is the point-discharge current through the point, F is the potential gradient at the ground nearby and M is the minimum value of F for discharge to occur.

This formula has been used by Whipple and Scrase (1936), Chiplonkar (1940), Hutchinson (1951) and Yriberry (1954); Chalmers (1952) has shown that such a relationship would be expected for a point which is one of a number in rectangular array, provided that wind is ignored and ions are removed solely by the electric field. However, Kirkman and Chalmers (1957) re-analysed the results to find the best value of n for fit with

$$I = B\,(F - M)^n \quad . \quad . \quad . \quad . \quad . \quad (2)$$

and found, for Whipple and Scrase's results, $n = 1{\cdot}07 \pm 0{\cdot}13$ and for Hutchinson's $n = 1{\cdot}09 \pm 0{\cdot}06$; in each case (2) with the best value of n, gave a better fit than (1).

* In order to distinguish between the different constants involved, they are here given different symbols from those used in the original papers.

For an isolated point, Chalmers and Mapleson (1955), using a tethered balloon found

$$I = CW^{\frac{1}{2}}(Fh)^{7/4} \quad . \quad . \quad . \quad . \quad . \quad (3)$$

where W is the wind-speed and h the height of the point. Kirkman and Chalmers (1957) found, for a point on a high mast

$$I = D(W + k)(F - M) \quad . \quad . \quad . \quad . \quad (4)$$

Chapman (1956) suggested, on semi-theoretical grounds, that the relation should be:

$$I = E(V - V_0)v \quad . \quad . \quad . \quad . \quad . \quad (5)$$

where V is the potential of the point relative to the surroundings, V_0 the minimum value of V for discharge and v the velocity with which the ions leave the point. For no wind, v is proportional to F and (5) does not differ much from (1), and for high wind, $v = W$ and (5) becomes (4) with k zero.

Chalmers (1957) suggested that v should be expressed as the vector sum of W and aV and this gave

$$I = G(V - V_0)(W^2 + a^2V^2)^{\frac{1}{2}} \quad . \quad . \quad . \quad (6)$$

so that aV represents the ion speed in the field. The results of Chalmers and Mapleson (1955) could fit (6) as well as (3), and those of Kirkman and Chalmers (1957) also fit (6) as well as (4). In addition, Large and Pierce (1957) applied a high potential to a point in the free air and found agreement with (6). But the values of the constants required to give good agreement are not the same for the different sets of observations.

3. Apparatus for artificial point

The point used was mounted on an insulator of polystyrene, supported at a height of about 15 m, extending about 2 m above a large horse-chestnut tree. The point was connected to a sensitive mirror galvanometer arranged to be critically damped.

The anemometer was of the three-cup generator type, giving alternating current which was rectified and measured by a mirror galvanometer. The anemometer was fixed to the same rod which held the point and was about 1 m below the point. The anemometer was calibrated in the wind tunnel of the National Coal Board at Longbenton, Newcastle upon Tyne.

The potential gradient was determined by an agrimeter (Chalmers, 1953) giving a current also measured with a sensitive mirror galvanometer, critically damped.

The three quantities were recorded on photographic paper on the same moving drum. In order to avoid waste of paper, a device was made to give records only when useful results were likely. For this purpose the current from the agrimeter was passed through a second, less sensitive, mirror galvanometer and the light beam from this galvanometer was focused on the sensitive area of a Mullard OCP 71 photo-transistor; it was arranged that when a potential gradient of $\pm$ 400 v/m existed the light beam moved off the sensitive area. Then the change of current in the photo-transistor operated a sensitive relay which, in turn, operated other relays; these switched on the recording-camera motor and galvanometer lamps and also a counter to denote the length of time for which the camera was operative.

The agrimeter was situated 15 m to the west of the tree carrying the point, and the situation of other trees and buildings were such that the current from the agrimeter would be unaffected by point-discharge space charge only if the wind was from a westerly direction. Therefore only those records were used in which this was the case, in contrast to the previous results where no attempt was made to avoid the effects of space charges, such as have been made evident by Maund and Chalmers (1960).

4. Results with artificial point

During the period from March 1958 to August 1960, there were 20 occurrences of potential gradient sufficiently high to give point discharge, but of these only 9 were suitable for analysis because of the requirement of a westerly wind; 5 of these gave negative currents (negative charge coming down) all at the same low wind-speed, and the other 4 positive currents at different wind-speeds. The highest potential gradient was 2,500 v m^{-1}, with a point-discharge current of 12 μa and a wind-speed of 5·5 m sec^{-1}.

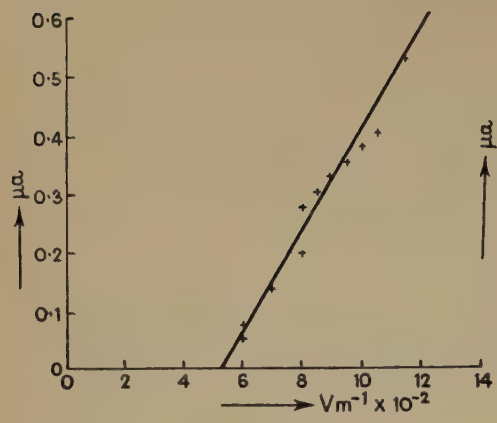

Figure 1. Negative point discharge and potential gradient (Wind-speed 3 m sec^{-1}).

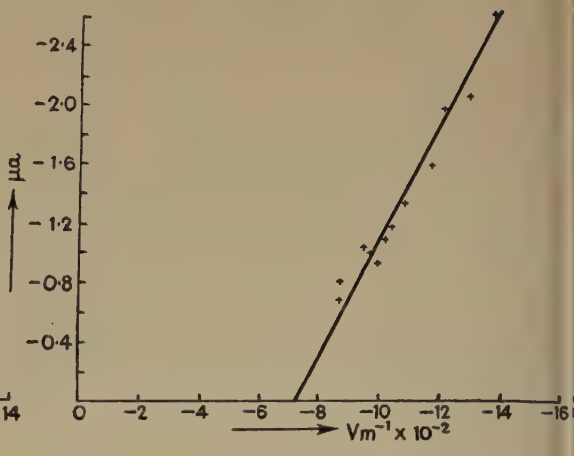

Figure 2. Positive point discharge and potential gradient (Wind-speed 3 m sec^{-1}).

For each particular wind-speed, the relation between point-discharge current and potential gradient was linear :

$$I = J(F - M)$$

Examples are shown in Figs. 1 and 2. Other things being equal, the negative point-discharge currents were greater than positive. When the wind-speed was very small, J was 0·0011 μa v^{-1} m for negative and 0·00073 μa v^{-1} m for positive currents. The mean values of M were $-$ 900 v m^{-1} and + 600 v m^{-1}.

The values of J for different wind-speeds for positive currents are shown in Fig. 3. While the number of results are insufficient to give very definite conclusions, agreement appears with

$$I = D(W + k)(F - M) \quad . \quad . \quad . \quad . \quad (4)$$

as found by Kirkman and Chalmers (1957).

If this is expressed in the form :

$$I = G(V - V_0)(W^2 + a^2 V^2)^{\frac{1}{2}} \quad . \quad . \quad . \quad . \quad (6)$$

with $V = hF$, $V_0 = hM$, h being the height of the point (15 m), the value of the constant G is found to be $6{\cdot}5 \times 10^{-11}$ and V_0 is 9×10^2 v; it is not possible to get any very certain result for the value of a^2 but an approximation is $2{\cdot}5 \times 10^{-9}$.

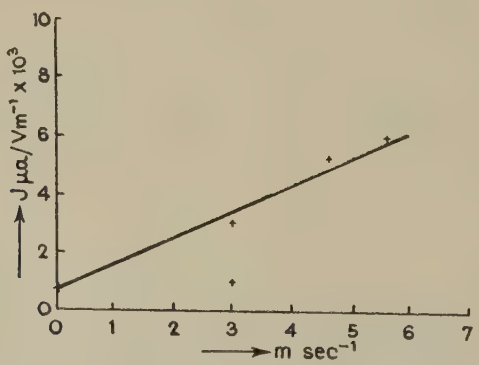

Figure 3. Variation of J with wind-speed (positive currents).

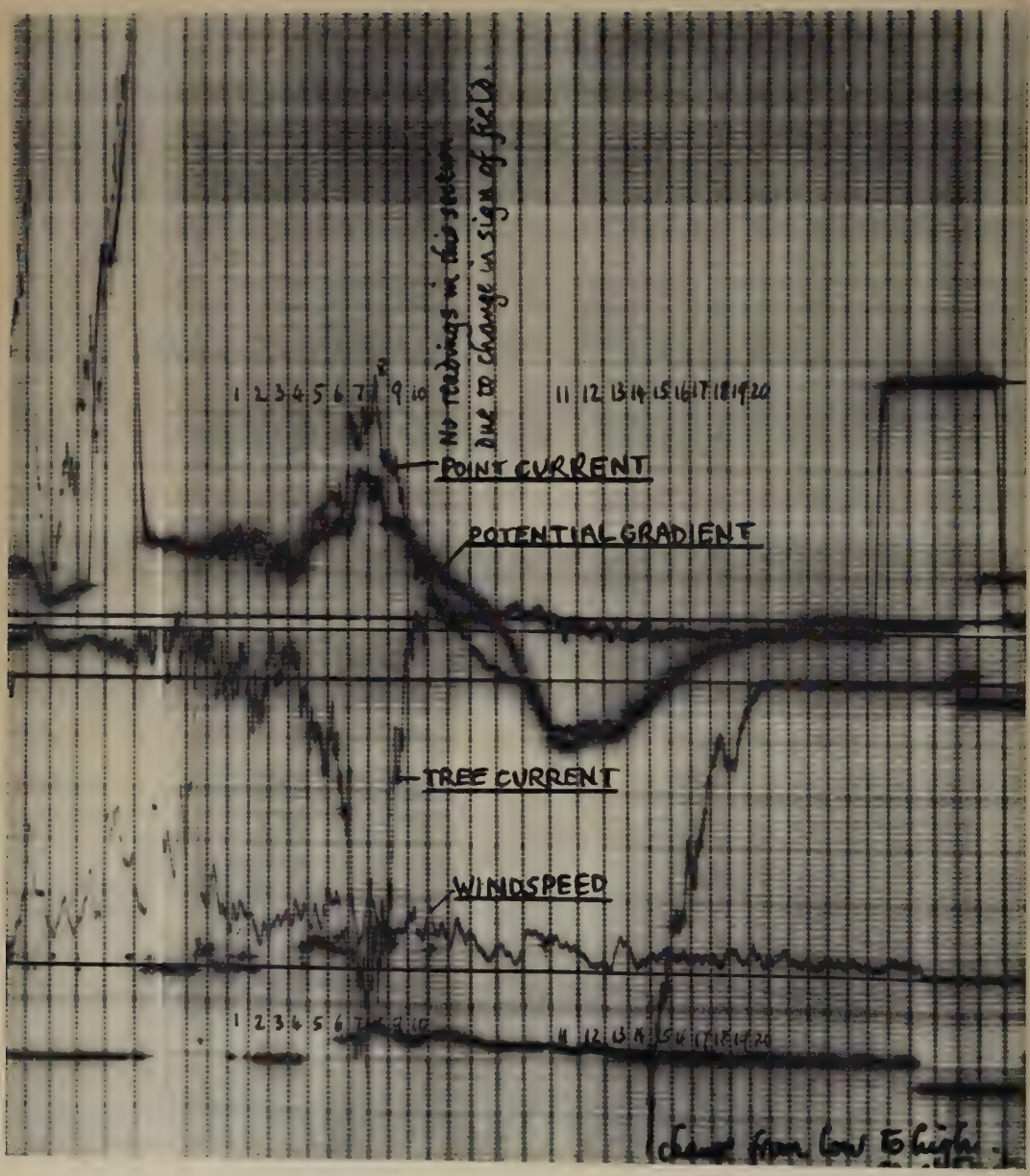

Figure 5. Record showing point discharge down the point and down the tree.

Using other values quoted and obtained by Chalmers (1957), the constants are :

	G ($\times 10^{-11}$)	V_0 ($\times 10^{-3}$)	a^2 ($\times 10^{-9}$)
Large and Pierce (1957)	1·6	7	8
Chalmers and Mapleson (1955)	6·5	2·5	80
Kirkman and Chalmers (1957)	60	8	2
Chapman (1956)	1·4		
Milner and Chalmers (1961)	6·5	9	2·5

It is not to be expected that G, V_0 and a should be the same for all types of point, but it is surely significant that the values for a point at tree-top height fall close to the values for points much higher.

5. Apparatus for tree

In order to measure the point-discharge current flowing down a tree, Kirkman (1956) inserted brass screws at two different levels and connected them to either side of a low-resistance galvanometer. He was not able to obtain useful results because of changes of zero, ascribed to contact potential effects. To avoid these the brass screws were replaced by mercury contacts; two small holes were bored through the bark of a lime tree, situated 20 m south of the tree which holds the artificial point and reaching to about the same height as the artificial point, and glass tubes were inserted containing mercury so as to give contact into the sapwood (Fig. 4); this proved more satisfactory than the brass screws; the two tubes were separated by a vertical distance of 3 m and the resistance of the tree between them was found to be about 4,000 ohms. Thus a galvanometer of resistance 5 ohms will take almost all the current flowing down the tree.

6. Results for tree

There has, so far, been only one occasion on which useful results have been obtained with the tree, and the record is reproduced in Fig. 5 (Pl. XIII); on other occasions of potential gradient, sufficiently large for point discharge, the wind direction was unsuitable. The relations between the tree current and the potential gradient, and between the current through the point at the same time and the potential gradient are shown in Fig. 6, with positive currents, and it is apparent that the same kind of linear law holds in both cases. The value of M, the potential gradient for onset of point discharge is about 1,000 v m^{-1} for the tree, as compared with about 300 v m^{-1} for the point under these conditions and 600 v m^{-1} on the average and the tree current is of the order of one-tenth to one-fifth of the point current. During the whole of the record, the wind was nearly constant at about 5·6 m sec^{-1}, so it was not possible to find how the current depends on the wind.

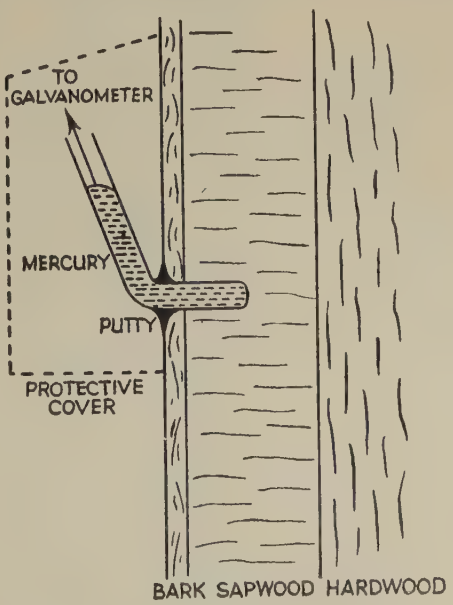

Figure 4. Tree electrode.

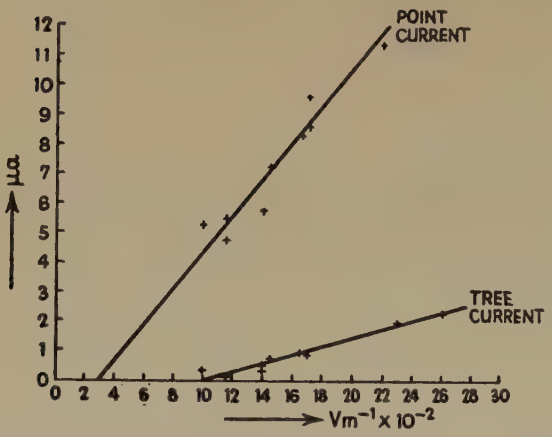

Figure 6. Simultaneous point and tree currents against potential gradient.

At the time of this record, the tree was in full leaf and the results may be compared with those of Maund and Chalmers (1960), who found, by an indirect method, that a tree without leaves gave currents of the same order as those from a point at the same height, while a sycamore tree in full leaf gave no detectable current.

It is intended to carry out further observations of the same kind.

7. Acknowledgments

It is a pleasure to acknowledge the assistance of members of the Atmospheric Physics Group at Durham. One of us (J.W.M.) is indebted to the D.S.I.R. for a maintenance grant.

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551.551.2 : 551.524.4

Similarity theory and temperature structure in the lower atmosphere

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(Manuscript received 20 February 1961)

SUMMARY

An equation based on similarity theory is used to explain the properties of various temperature statistics. Extensive observations indicate that wind and temperature profiles are nearly similar, and the ratio of the exchange coefficients for heat and momentum is of order one.

1. VERTICAL VARIATION OF STANDARD DEVIATION OF TEMPERATURE

Similarity theory suggests that the standard deviation of temperature, σ_t, should vary with the minus one-third power of the height z above the ground in the case of free convection, and that it should be independent of height under near-neutral conditions.

Priestley (1960) evaluated the power p in the relation

$$\sigma_t = Bz^p \qquad (1)$$

from several sources, and plotted it as a function of Richardson number : 600 individual values are represented by the points. The observations of Priestley's paper are shown in Fig. 1. As Priestley has pointed out, the data confirm the limiting powers : an attempt will now be made to account for the intermediate powers.

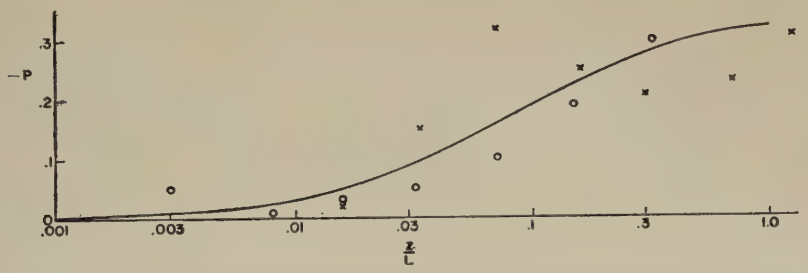

Figure 1. Power in vertical distribution of standard deviation of temperature as function of z/L. Circles : from Iowa State College. Crosses : from M.I.T. and C.S.I.R.O. (after Priestley).

Let θ be the potential temperature, and l, a mixing length. If the pressure in near 1,000 mb, we can put :

$$T' = l\frac{\partial\theta}{\partial z} \qquad (2)$$

Here, T' is the deviation of temperature from its mean. Taking the r.m.s. on both sides, we find :

$$\sigma_T = \sqrt{\overline{l^2}}\,\frac{\partial\theta}{\partial z} \qquad (3)$$

(a bar indicates the mean).

The quantity $\sqrt{\overline{l^2}}$ is a measure of the average distance travelled by the eddying air, which, in neutral and unstable conditions, can be put proportional to height. We therefore find :

$$\sigma_T = Az\frac{\partial\theta}{\partial z} \qquad (4)$$

where A is a constant. Priestley confirms Eq. (4) from his observations, with A about 1·3.

In a stably stratified atmosphere, the quantity $\sqrt{\overline{l^2}}$ would be expected to be less than the height. Thus the ratio, $(\partial\theta/\partial\,(lnz))/\sigma_T$ would be expected to be constant in unstable and neutral air, but increase with increasing stability in stable air. Observations by Zwang (1960) confirm this result.

In order to proceed with the argument, we will assume that the profiles of wind and potential temperature are similar. This assumption is quite controversial. It is at variance for example, with conclusions reached by Ellison (1957). However, it is consistent with similarity theory, and some recent accurate observations by Sheppard (unpublished) also confirm similarity of profiles.

Additional evidence for near-similarity of potential temperature and wind profiles can be deduced from the extensive records taken during Project Prairie Grass at O'Neill, Nebraska, during 1956, under the direction of M. H. Halstead (see Covey *et al.* 1958).

First, curvature parameters for the profiles were defined by :

$$C_t = \frac{T_8 - T_2 \quad + 0{\cdot}06°\mathrm{C}}{T_1 - T_{0{\cdot}25} + 0{\cdot}01°\mathrm{C}} \text{ and } C_v = \frac{V_8 - V_2}{V_1 - V_{0{\cdot}25}} \qquad . \quad . \quad (5)$$

where T is temperature, V is wind speed, and subscripts indicate the height of the observation in meters. If the profiles are similar, C_t and C_v should be equal. Table 1 lists the ratios of C_v/C_t as a function of the Richardson number between 2 m and 8 m.

TABLE 1. TEST OF PROFILE SIMILARITY

Ri_{2-8}	$Ri>0{\cdot}1$	$0{\cdot}1>Ri>0$	$0>Ri>-0{\cdot}1$	$-0{\cdot}1>Ri>-0{\cdot}4$	$-0{\cdot}8>Ri>-0{\cdot}4$	$Ri>-0{\cdot}8$
C_t/C_v	0·88	0·99	0·83	0·82	0·87	0·97
σ	0·32	0·26	0·26	0·30	0·22	0·32
N	26	69	45	30	15	7

Here, N is the number of situations included in each column. σ measures the standard deviation in each Richardson number category. It is not the standard deviation of the mean, since successive observations are often included and the usual formula for standard deviations from means assumes independence of the various statistics from each other.

Table 1 shows that the ratio of the curvature parameters is always near unity. The ratio is less than one; whether this result is significant is uncertain, particularly since the deviation from unity is not related to the Richardson number.

From Table 1, some tentative conclusions can be drawn concerning the relative behaviour of the exchange coefficients for heat and momentum. Under the usual assumption that heat flux and stress are constant with height in the lowest 10 m, we have :

$$C_t = \frac{\int_2^8 \frac{dz}{K_h}}{\int_{0{\cdot}25}^1 \frac{dz}{K_h}} \text{ and } C_v = \frac{\int_2^8 \frac{dz}{K_m}}{\int_{0{\cdot}25}^1 \frac{dz}{K_m}} \qquad . \quad . \quad . \quad (6)$$

where K_h and K_m are the exchange coefficients for heat and momentum, respectively. Now, let the layer from 2 m to 8 m be designated by subscript t, and the layer from 25 cm to 1 m by subscript b. Then :

$$C_t/C_v = \frac{K_{hb}}{K_{mb}} \Big/ \frac{K_{ht}}{K_{mt}} \qquad . \quad . \quad . \quad . \quad (7)$$

Here, the exchange coefficients are means in the layers indicated by the subscripts. If the ratio of the curvature parameters was exactly unity, the ratio of exchange coefficients for momentum and heat would be independent of height. It is frequently assumed that this ratio is near unity for low Richardson numbers (low heights). It then follows that the exchange coefficients are essentially equal at all heights. If we take the deviation of the ratio C_t/C_v from unity as real, it would indicate that K_h exceeds K_m by perhaps 10 per cent more in the top layer than in the bottom layer. In any case, the ratio of the coefficients is likely to be of order one at all Richardson numbers.

We now proceed with the original argument. A non-dimensional wind shear, S, is defined by :

$$S = k\sqrt{\frac{\rho}{\tau}}\, z\, \frac{\partial V}{\partial z} \quad . \quad . \quad . \quad . \quad . \quad (8)$$

Here, τ is the surface stress, ρ the density, and k von Karman's constant. Under the assumption of profile similarity, the quantity S is proportional to $z\,\partial\theta/\partial z$ and therefore, according to Eq. (4), to σ_T.

According to Monin and Obukhov (1954), S is a universal function of z/L, where L is defined by $\tau^{3/2} c_p\, T/\rho^{1/2}\, kg\, H$ (g is gravity, c_p is the specific heat at constant pressure, H the vertical flux).

According to Yamamoto (1959) and Panofsky, Blackadar and McVehil (1960), S as a function of z/L is given by :

$$S^4 + \frac{\gamma z}{L} S^3 = 1 \quad . \quad . \quad . \quad . \quad . \quad (9)$$

According to the latter paper, Eq. (9) fits observations well with the parameter γ equal to 14.

By logarithmic differentiation of Eq. (9), we find :

$$p = \frac{\partial \ln \sigma_T}{\partial \ln z} = \frac{\partial \ln S}{\partial \ln z/L} = \frac{\gamma z/L}{4S + 3\gamma\, z/L} \quad . \quad . \quad . \quad (10)$$

Since S is a function of z/L (Eq. 9), Eq. (10) gives p as function of z/L alone.

Figure 1 shows Eq. (10) fitted to the observations as evaluated by Priestley. Apparently, the fit is satisfactory. This good agreement is not intended to imply that Eq. 9 is the only equation able to account for the behaviour of p. Other formulae which interpolate between free convection and forced convection are likely to be equally consistent with the observations, provided that they contain an adjustable parameter like γ, particularly in view of the uncertainty of the data. As might have been anticipated, the same choice of γ which is able to account for the transition from forced to free convection in the case of the wind profile also explains the characteristics of p.

The physical meaning of the parameter γ is the relative efficiency of mechanical turbulence and convection for transferring heat or momentum. Since convection eddies are larger than mechanical eddies (see, e.g., Gurvic 1960) convection mixes air with different characteristics more efficiently than mechanical turbulence, and produces relatively greater flux. As a result, when Richardson number is minus one, so that mechanical energy and convection are produced at the same rate, heat convection prevails and the temperature profile and the fluctuations of wind and temperature have the characteristics of ' free ' convection. The transition from forced to free convection occurs at much lower Richardson numbers for the same reason. Of course this interpretation is quite tentative.

2. The relation between z/L and Richardson number

In many instances, the vertical heat flux is not measured, so that the quantity z/L is not given. However, temperature and wind profiles are often available, so that the Richardson number, Ri, is given. For that reason, Priestley has prepared a diagram which shows the empirical relationship between Ri and z/L. Priestley's diagram is here reproduced as Fig. 2.

With the aid of Eq. (9), a theoretical relationship between Ri and z/L can be derived. By definition of the quantities involved, we have :

$$\frac{z}{L} = -\,SRf \quad . \quad . \quad . \quad . \quad . \quad . \quad (11)$$

where Rf is the flux Richardson number. Elimination of S between Eqs. (9) and (11) gives

$$\frac{z}{L} = -\,\frac{Rf}{4\sqrt{1 - \gamma Rf}} \quad . \quad . \quad . \quad . \quad . \quad (12)$$

Since Rf equals $Ri\, K_h/K_m$, we have finally :

$$\frac{z}{L} = -\,\frac{K_h}{K_m}\,\frac{Ri}{4\sqrt{1 - \gamma\, K_h/K_m Ri}} \quad . \quad . \quad . \quad (13)$$

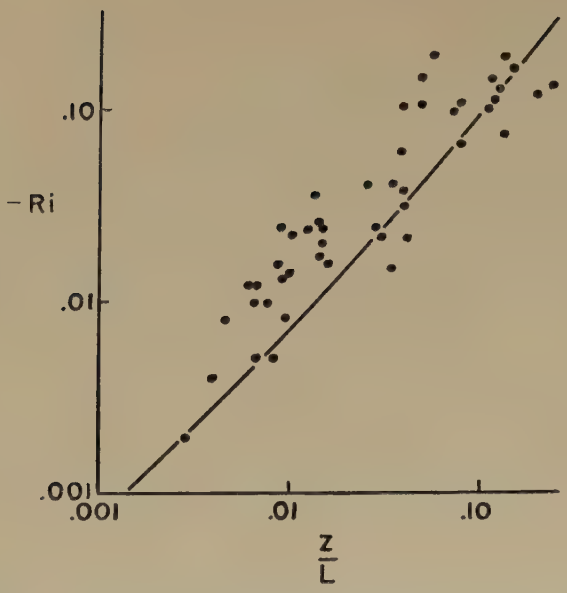

Figure 2. Relation between Richardson number and z/L (after Priestley).

According to the paper by Panofsky, Blackadar and McVehil (1960), the ratio of the exchange coefficients best suited to the observations quoted in that paper was 1·3. Eq. (13) with this ratio of the coefficients is shown in Fig. 2. The fit to the observations is fair, and would have been improved if the turbulent Prandtl number had been assumed to be unity.

In this discussion, the effect of moisture on Ri and L has been neglected. Actually, both these quantities refer to virtual temperature, and this omission may be serious over a wet surface. However, at O'Neill, differences between the lapse rates of temperature and virtual temperature are not important.

3. Heat flux and Richardson number

Perepelkina (1959) has determined the vertical heat flux from fluctuations of temperature and vertical velocity. She has also given the ratio: $-H/k^2 z^2 (\partial V/\partial z)(\partial\theta/\partial z) c_p \rho$ as a function of Richardson number. The result is given in Fig. 3, where each point represents the average of many observations.

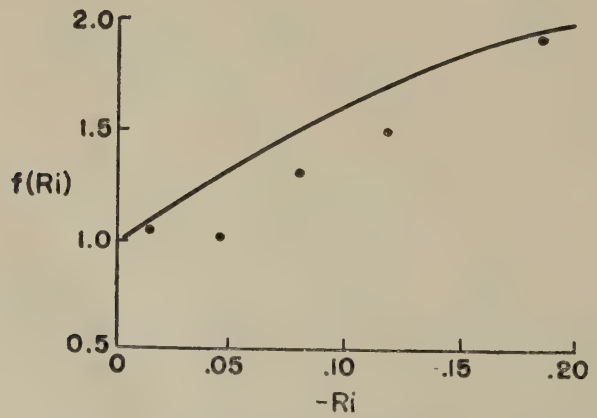

Figure 3. Ratio of heat flux and heat flux expected in neutral stability, as function of Richardson number (after Perepelkina).

The theory for this relationship can again be found from Eq. (9). On elimination of z/L between this and Eq. (11), we obtain

$$S = kz\sqrt{\frac{\rho}{\tau}}\frac{\partial V}{\partial z} = \frac{1}{4\sqrt{1 - \gamma K_h/K_m Ri}} \quad . \quad . \quad . \quad (14)$$

An equation very much like Eq. (14) was first given by Kazansky and Monin (1956) and also by Ellison (1957). K_m is defined by :

$$K_m = \tau \Big/ \left(\rho \frac{\partial V}{\partial z}\right) \quad . \quad . \quad . \quad . \quad . \quad . \quad (15)$$

Elimination of stress between Eqs. (14) and (15) yields :

$$K_m = k^2 z^2 \frac{\partial V}{\partial z}\sqrt{1 - \frac{K_h}{K_m} Ri} \quad . \quad . \quad . \quad . \quad (16)$$

and thus :

$$- H/\rho c_p k^2 z^2 \frac{\partial V}{\partial z}\frac{\partial \theta}{\partial z} = \frac{K_h}{K_m}\sqrt{1 - \frac{K_h}{K_m}\gamma Ri} \quad . \quad . \quad . \quad (17)$$

Fig. 3 compares Eq. (17) with observations, with $\gamma = 14$ as before, but with the ratio of the exchange coefficients equal to one. Apparently, the fit is reasonably good. If the ratio of the coefficients had been taken as 1·3, the agreement would have been less good, with theory systematically over-estimating the ratios. However, it is quite possible that the measured heat fluxes are too small, because the heat fluxes were determined only from 100-sec averages.

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551.521.16

The standardization of daylight measurements

By R. H. Collingbourne, P. Petherbridge and G. Spurr

Data on the availability of daylight at particular localities are required for such purposes as assessing window sizes and positions needed to ensure adequate daylighting in new buildings, estimating the lighting loads likely to be met by electricity generating equipment at different times of the day throughout the year and correlating plant growth with degree of exposure to the sky. For architectural design purposes in particular, these data are being increasingly required for many different parts of the world, and hence have to relate to a wide range of climatological conditions. Furthermore, as daylighting design techniques become more refined, the associated data need to extend beyond the basic measurement of sky illumination on an unobstructed horizontal plane outdoors.

Whereas measurements of the total radiation from the sun and sky at the earth's surface have been carried out at meteorological stations and elsewhere as a regular routine for a considerable number of years, it is only comparatively recently that any extensive measurements of visible radiation have been made and then only at a limited number of recording stations. Consequently, whereas an extensive body of knowledge on the measuring techniques employed for total radiation has been accumulated and has found its way in the form of guidance or recommendations into manuals on meteorological instruments and measurements, corresponding information on daylight measuring techniques is not by any means so well documented.

The Commission Internationale de l'Eclairage (CIE) is interested in the question of daylight measurements, and having prepared a report on the climatological data needed for natural lighting problems, is currently compiling recommendations on the measuring techniques to be used in the recording of such data. For this country, a need has been felt for at least some degree of standardization of daylight measurements, both in the types of measurement that are made and, to a lesser extent, in the instrumental techniques used for obtaining these measurements and in the presentation of the resultant data. A report on the subject has been compiled by the present writers for the Daylight Sub-committee of the National Illumination Committee of Great Britain, and an article based on this report has now been published (*Light and Lighting*, **54**, 1961, p. 71).

The report offers the following types of daylight measurement for which some degree of standardization would be required :

(*i*) Total (sun and sky) and diffuse (sky) illuminations on an unobstructed horizontal surface

(*ii*) Direct solar illumination.

(*iii*) Luminous efficiency of natural daylight.

(*iv*) Sky luminance distribution.

(*v*) Spectral composition of sun and sky radiation.

No specific recommendations are made for this country in respect of measurements of vertical surface illumination.

Appropriate recordings should be made so that measurements (*i*) to (*v*) can, if required, be correlated with cloud coverage (type, amount, height of cloud base and thickness), duration of sunshine and sun position (particularly solar altitude).

Another section of the report deals with techniques for daylight measurement, referring in particular to the colour and ' cosine ' responses of the light receptors and their sensitivity to temperature and humidity changes, the calibration of the measuring equipment and present-day techniques for data recording.

A final section of the report is concerned with methods for presenting illumination data, luminous efficiency values and sky luminance distributions. The photometric units, metric or non-metric, to be adopted will depend on the use to be made of the data.

Comments on the proposals are invited by the compilers of the report, R. H. Collingbourne (Kew Observatory), P. Petherbridge and G. Spurr (Central Electricity Research Laboratories).

Kew Observatory, Richmond, Surrey.
D.S.I.R., Building Research Station, Garston, Watford, Herts.
Central Electricity Research Laboratory, Leatherhead, Surrey.
2 June 1961.

551.510.534 : 551.513.1 : 551.524.7

Total ozone and the middle stratosphere over arctic and sub-arctic areas in winter and spring

By W. L. GODSON

(Read 15 March 1961. Presented by Dr. A. W. Brewer. See *Q.J.*, **86**, p. 301)

Mr. J. S. Sawyer : What horizontal structure are we to envisage for the baroclinic waves in the stratosphere to which Dr. Godson refers ? If these were in any way similar to baroclinic disturbances in the troposphere one would not expect them to give rise to periods of the order of 20 days since they would move with a speed comparable with that of the basic current, say 25 to 50 m sec^{-1}. Since the stratospheric jet stream forms a more or less distorted ring which is rarely centred at the pole, may not the fluctuations in temperature and ozone be as readily described in terms of the displacement and expansion or contraction of this ring as in terms of ' waves ' ?

Mr. H. H. Lamb : I wonder whether the behaviour of the stratospheric vortex over the Antarctic during the spring warming was really so different from that over the Arctic, as had been made out. I have analysed daily charts for the 50 mb level over the Antarctic during the spring warming period from early October to mid-November 1959. On the Australian side, the course of the warming, as in the Arctic, was a history of abrupt rises of temperature (by 10 to 20°C) followed by reversals in which the temperature returned, at least briefly, to the former levels; and this was several times repeated. So in this sense there were waves there too. Looked at on the map these changes appeared as migrations of the eccentric oval, or ' egg-shaped,' vortex from one sector to another, generally eastward (at least in the 1959 sequence), but the Pole and the Weddell Sea (Halley Bay) remained within the cold vortex throughout all the earlier migrations. The temperature rise at the South Pole was smooth. Possibly Halley Bay has appeared (in the few years for which observations are available) to experience only one quick, and final, warming just because that sector tends to be the last reached by warmer air. The warm air at length arrives there when the winter stratospheric vortex is finally dislodged from the polar region, migrates towards the Atlantic and decays.

Mr. E. Gold : I should like to ask two questions. If I have understood rightly, the increase in ozone over the Arctic is due to advection – horizontal transfer. What do the winds show ? Not the resultant winds deduced from the mean monthly or seasonal isobars but the winds actually observed or deduced from the uconals. One might expect these to have, on occasions, meridional components strong enough to transfer the air across the Arctic in hours or days rather than weeks or months.

My second question is to ask about the difference between the sequence following the winter increase in 1958/9 and that of the other years. In the former, the increase was maintained at roughly its initial level, whilst in the other years it soon fell appreciably from its initial level.

Dr. A. W. Brewer (*in reply*) : The impression which I have gained recently is that the northern hemispheric stratosphere is trying to do what the southern stratosphere does, but that it is not able to achieve it, presumably owing to the much larger disturbance to the flow caused by the irregular distribution of land and sea in the Northern Hemisphere.

Dr. W. L. Godson (*in reply, communicated*) : In reply to Mr. Sawyer, middle-stratospheric baroclinic waves are characterized by larger values of period and wavelength and smaller values of wave-speed than those in the troposphere and lower stratosphere. Advection and vertical motion are much more nearly in balance than at lower levels, and advection generally predominates, but certainly not universally. Ridges are warm and troughs cold, and occasionally migrant cyclonic and anticyclonic cells split off – not at all unlike tropospheric circulations. Temperature and ozone anomalies progress in a relatively regular manner, almost invariably from west to east, and have a distinct and coherent wave structure. Independent evidence is available from the many synoptic charts and analyses for levels up to 10 mb published in the last two or three years.

In reply to Mr. Gold, as referred to in my reply to Mr. Sawyer, middle stratosphere flow in winter is characterized by a very broad belt of relatively strong westerlies about an intense polar vortex, centred not too far from the pole. Baroclinic waves are imbedded in the flow and they (including the strong temperature and ozone anomalies associated with them) move at very much less than the wind speeds, since vertical motions almost exactly compensate horizontal advective effects (certainly for temperature and presumably so for ozone in general).

For late final-warming winters (such as 1958/59), the final warming (and ozone maximum) marks the permanent transition from westerly to easterly flow in the middle stratosphere and thus the end of vigorous dynamic activity. Subsequently, photochemical decay is probably a large factor in the slow decrease of mean total ozone through the summer. For early final-warming winters (such as 1956/57 and 1957/58) the final-warming wave and its associated ozone maximum is generally more intense then in the other case, and this is borne out by extremely high temperatures in the 10-30 mb layer. The polar vortex is not destroyed, but is pushed well off the pole and weakened. Subsequently it migrates poleward, and westerly flow is re-established. During this process temperature and ozone minima are produced locally and, in addition, the excess over photochemical equilibrium values is much greater than for the later final-warming cases and destruction of ozone is accelerated, relatively. It is interesting to note that the strongest peak in ozone maxima at Edmonton occurred in early 1952, when the intense final-warming documented by Scrase for Lerwick moved westward across North America.

In reply to Mr. Lamb, the chief difference between the winter middle stratosphere over the Arctic and Antarctic is that the initiation of the final-warming wave (which is *not* followed by a temperature reversal to former levels) over the Arctic has never been later than the end of March (based on almost 25 years of temperature and ozone data). Antarctic and sub-Antarctic data (back to 1948) suggest no initiation of the final warming prior to October (April in the Northern Hemisphere), and certainly *no* cases of an early final warming. Furthermore, strong warm-belt activity and an intense stratospheric jet stream commence much earlier in the Arctic regions. Maximum ozone amounts at British Antarctic stations during the final warming are very much less than those for Arctic stations which should be roughly comparable. The multiple final-warming (or stepwise approach of summer stratospheric temperatures) is relatively uncommon in the Arctic, where, very frequently, winter minimum temperatures immediately precede the final-warming wave. It is not unusual in the Arctic to find strong baroclinic waves in November or December that produce, even at high latitudes, temperatures in the middle stratosphere comparable to those of early September. It is most unlikely that the Antarctic stratosphere has had or will have corresponding events.

551.501.81 : 551.578.71

Radar scatter by large hail

By D. ATLAS, W. G. HARPER, F. H. LUDLAM and W. C. MACKLIN

(Read 17 May 1961. Presented by Mr. W. G. Harper. See *Q.J.*, **86**, p. 468)

Dr. J. M. Stagg (*President*) : Your description of the physics of the process of scattering and attenuation of the radar beam, that there is interference between the part reflected from the front surface of the sphere and the part that enters the sphere and is reflected from within it, seems to account well enough for the oscillations in the curve of back-scatter cross-section against D/λ in the case of ice spheres. But for water spheres there is little internal reflection because the absorption is great, and yet the curve for water is as oscillatory as for ice. Is there any simple explanation ?

Professor B. J. Mason : Did Dr. Ryde work from the exact Mie theory in computing his back-scattering curve for ice ?

Mr. W. G. Harper (*in reply*) : Dr. Stagg has touched on a difficult point. Oscillations similar to those of the water curve occur in the case of perfectly conducting spheres where there is no internal penetration by the beam. They are caused by a surface wave propagated around the back surface which interferes with the reflection from the front surface. The water sphere curve behaves similarly because of its high absorption. I know that Mr. Probert-Jones has done some work on this, and I wonder if he would care to add anything to what I have said.

In reply to Professor Mason, the sizes of particles Ryde was dealing with were not small enough for the Rayleigh approximation to hold, and there is no doubt that he used the complete Mie theory. It was a remarkable achievement, without an electronic computer, for him to have obtained the results he did.

Mr. J. R. Probert-Jones : There is no doubt that the authors are correct in attributing the high level of the back-scattering cross-section of large ice spheres to a ray reflected off the back surface. A simple application of Snell's law shows that a ray incident at an angle of 56° to the surface will be so reflected and emerge parallel to the incident direction. The theory of geometrical optics gives an infinite intensity for this ray ; in fact the assumption that the emergent wave-front is plane is not satisfied. It is of interest to note that this ray only exists for spheres whose refractive indices lie between $\sqrt{2}$ and 2. Moreover, it is this ray, with the addition of a small section of surface wave, which gives rise to the phenomenon of the ' glory ' in water drops at optical wavelengths. It is convenient therefore to refer to it as the ' glory ray.'

(Discussion 6 June 1961 at Toronto Conference)

Prof. W. Hitschfeld : One is greatly impressed with the ingenuity which the authors displayed in arranging for a field experiment using a straightforward approach which yielded such a rich and theoretically satisfactory body of data. I would like to comment, however, on the meteorological significance of these results. In Alberta, over the last few summers, Dr. Douglas has collected hail on 28 occasions, and was able to reconstruct the size-distribution in space, as reported in a recent letter to *Nature*. Then, using the results obtained by the present authors, he calculated the radar reflectivity of this hail, on the assumption that it was either wet or dry. In 26 of these cases it was found that the contribution from the smaller stones predominated to such an extent that the wet assumption would actually produce a stronger signal than the dry assumption. This is easily understood by the authors' Fig. 5, according to which, water scatters more strongly than ice for diameter/wavelength ratios less than 0·9. Obviously for a radar wavelength of 3·3 cm, much of the hail is below this limit, and *a fortiori* even more so at longer radar wavelengths.

I feel this finding is important : whilst it is conceivably not generally applicable to hail everywhere, it might serve as a warning against using the strength of the radar echo as a criterion of the wetness or dryness of the hail.

Dr. D. Atlas (*in reply*) : The fact that in only 2 out of 28 Alberta distributions would the wet hail produce a stronger signal than the dry is undoubtedly due to the modest size of most Alberta hail*. The occurrence of a reflectivity decrease upon wetting is to be expected only when the hail diameter is about equal to, or larger than, the wavelength, and so the interesting points in Douglas's data were the two exceptions – which were undoubtedly associated with the largest hail. At 3 cm wavelength, a maximum aloft in the vertical reflectivity profile, such as those reported by Donaldson in New England, would therefore be useful in identifying the presence of stones larger than 3 cm diameter.

* This remark hurt the pride of native Albertans present and was answered with a photograph of a 4-in. diameter stone.

551.578.7

The fallspeeds of hailstones

By W. C. MACKLIN and F. H. LUDLAM

(Read 17 May 1961. Presented by Dr. F. H. Ludlam. See *Q.J.*, **87**, p. 72)

Dr. J. M. Stagg (*President*) : In this paper you have hailstones falling predominantly with their shortest axis vertical : but at some other times you have left me with a picture of hailstones growing by the water from the downward-facing surface being swept round to the back and so piling up and freezing on the upward-facing surface. If this is the mechanism of growth the longest axis should be vertical. Which of the two have I got wrong ?

Dr. F. J. Scrase : As the drag coefficient depends on the air density the measurements of the coefficient in the wind tunnel and of fallspeeds near the ground may not apply to the fall of hailstones at high levels.

Mr. J. B. Andrews : Has Dr. Ludlam found evidence so far in his observations to the attainment by hailstones of the second terminal velocity, which previous workers have suggested might be reached by large hailstones ?

Dr. F. H. Ludlam : In reply to Dr. Scrase, the drag coefficient is a function only of Reynolds number and the relation can be derived from measurements at any air density. It is true that in the high troposphere the air density and the drag on a given stone are less, and its fallspeed there is about half as great again as near the ground, a difference which needs to be taken into account in any theory of hailstone growth. We have not so far observed or experimented upon stones large enough to have Reynolds numbers in the region where a sudden drop in drag coefficient occurs and the substantially greater terminal speeds (as mentioned by Mr. Andrews) can arise.

In reply to Dr. Stagg, I think it is true that excess liquid collected by a stone is swept to its upper surface, but there the rate of accumulation of ice is probably less than on the lower surface. I must admit that I do not know just what determines the form of a growing stone, but I am fairly confident that at least in free fall, after growth, a stone which is oblate will have the attitude described.

551.501.81 : 551.574.13 : 551.578.71

Calculations of Mie back-scattering of microwaves from ice spheres

By B. M. HERMAN and L. J. BATTAN

(Read 17 May 1961. Presented by Mr. W. G. Harper. See *Q.J.*, **87**, p. 223)

Mr. W. G. Harper : I am very glad to have had the opportunity of presenting this paper by Mr. Herman and Dr. Battan which so fully confirms and greatly extends the results that have been obtained in this country (see *Q.J.*, **86**, p. 468).

I fully agree with their conclusion that very low absorption is the main factor in causing ice spheres to have such large back-scattering cross-sections, but I feel I must question their explanation of the transition shown in their Fig. 5, namely that until the critical value for absorption is reached the quantity of energy penetrating the drop and focusing at the back surface is virtually zero, but that at this value the process suddenly becomes of importance. The exponential decay due to absorption in water at 3·2 cm wavelength is however such that the power density of radiation has fallen to 10 per cent of its initial value in a penetration of about 2 mm. The authors reduced the absorption coefficient by a factor of 10, at which value the penetration would be 2 cm before reduction to the 10 per cent level. At an α equal to 6 the path length through the dielectric would be nearly 12 cm, and the power density would have dropped to one millionth. Clearly this is still a very strong absorption at these size spheres, and it could no longer affect the shape of the curve. The actual absorption coefficient for ice lies between 1/100 and 1/1,000 of that for water, depending on wavelength and temperature, and it would have been of interest if they had carried their reduction of absorption further. Another factor relevant to this has been made by Mr. Probert-Jones in the discussion of the ' Radar scatter by large hail ' (see discussion p. 604), namely that the internally reflected ray is only possible at values of refractive index between $\sqrt{2}$ and 2 (van de Hults 1957). The refractive index of 8·99 used in computing Fig. 5 was well outside this range.

Reference

Van de Hulst, H. C. 1957 *Light scattering by small particles* John Wiley & Sons, New York.

Dr. Davé (*communicated*) : Is it possible that the longer-period oscillation in the ice curve (Fig. 1, p. 225), with a wavelength of about 7 in α, could have been caused by truncation errors in computation ?

Mr. D. C. Thomas : Dr. Ludlam when presenting the paper on ' Fallspeeds of hailstones,' quoted field observations which show that very large hailstones are seldom true spheres, and the paper gives results for the fallspeeds of oblate spheroids as well as for true spheres. The values of reflectivity of hailstones in the two papers presented by Mr. Harper refer only to spheres. It is possible that an oblate ice spheroid has a higher reflectivity than an ice sphere for certain attitudes to the beam of radiation, thus giving even greater reflectivity relative to the equivalent water sphere. Has Mr. Harper any results either theoretical or practical for the reflectivity of oblate spheroids ?

Mr. W. G. Harper (*in reply*) : We have made some measurements recently at Malvern with perspex spheroids which seem to show what Mr. Thomas has suggested. Perspex has a rather similar dielectric constant to ice. We have not tried to work with ice spheroids. I do not think that these large spheroids are amenable to theoretical treatment.

Mr. B. M. Herman and Dr. L. J. Battan (*in reply, communicated*) : In reply to Mr. Harper's comments concerning our interpretation of Fig. 5, we agree that the physics of the back-scattering process is considerably more complicated than the rather elementary explanation presented in our paper. It is generally accepted that low absorption is a necessary condition for the rather anomalous behaviour of σ_b with increasing α, such as is the case of ice. Our Fig. 5 shows that a transition in the character of the σ_b curve may occur for a relatively small change of the absorption coefficient. About all that may be said at this point is that this transition is somehow associated with the small reduction in the absorption coefficient. With the value of m ($= 8{\cdot}99$) used in Fig. 5, it is impossible to have any rays come out parallel to themselves, other than the axially incident ray, as pointed out by Mr. Harper. Therefore, interference between the axial ray reflected at the front surface of the sphere, and that reflected at the back surface is probably of considerable importance in this case. Obviously, reflection from the back surface will be of greater importance for spheres with small absorption. A value of $k = 0{\cdot}009$, as used in Fig. 5, implies an exponential absorption coefficient of approximately 0·32. For an $\alpha = 6{\cdot}0$, the axial ray must traverse a path length of about 12 cm through the sphere, and would thus be reduced to about 2 per cent of its original value by absorption alone. Since only a part of the ray is reflected at the back surface, this value would be further reduced. Thus, it is evident that there must be other effects of importance in this problem. In the case of ice, the focusing effect does occur, and therefore there are other rays to consider in addition to the axial rays. Dr. D. Atlas has recently pointed out to us an article by Mr. Probert-Jones in which it is demonstrated that many of the oscillations in the curve of σ_b for ice may be accounted for by surface waves progressing around the sphere. This effect may also be of importance for those curves presented in Fig. 5.

In reply to Dr. Davé, it is unlikely that the oscillations referred to are due to truncation errors. The summations in the Mie equations were carried out until the last term of the series caused no change in the value of σ_b to eight digits. Since the peaks of these oscillations occur at values of diameter, $D \approx n\lambda$, it is likely that they are due to interference effects, probably between the axially reflected rays. However, an exact interpretation of the many facets of the σ_b curve for ice must await until such a time as a more complete understanding of the entire back-scattering mechanism is obtained.

551.510.53 : 551.511.3 : 551.557.33

Quasi-periodic wind variations with height in the lower stratosphere

By J. S. SAWYER

(Read 21 June 1961. See *Q.J.*, **87**, p. 24)

Dr. R. J. Murgatroyd : The fluctuations of wind as presented by Mr. Sawyer show a marked resemblance to the detailed profiles of ozone amounts obtained in the lower stratosphere using the chemical type of radiosondes. In both cases there appears to be marked stratification. Would any connexion of this type be expected ?

I was interested in the statement that the readings on descent seem to be affected by a lag in measurements obtained when the parachute is used. Is there any more information on this point ?

Mr. B. C. V. Oddie : I have examined many of the records of the radar theodolite, and the regularity and exactness with which the features shown during ascent are reproduced during descent is most striking – they have to be seen to be appreciated. The small lag shown in descent, as compared with ascent, is probably due, at least in part, to the lag in the response of the radar theodolite : a lag deliberately incorporated into the design for the sake of stability.

Mr. Sawyer's explanation of these variations of velocity was very clear and convincing in its mathematical form, but I still have difficulty in forming a physical picture of the movements of the air. Can Mr. Sawyer describe the flow in general terms ?

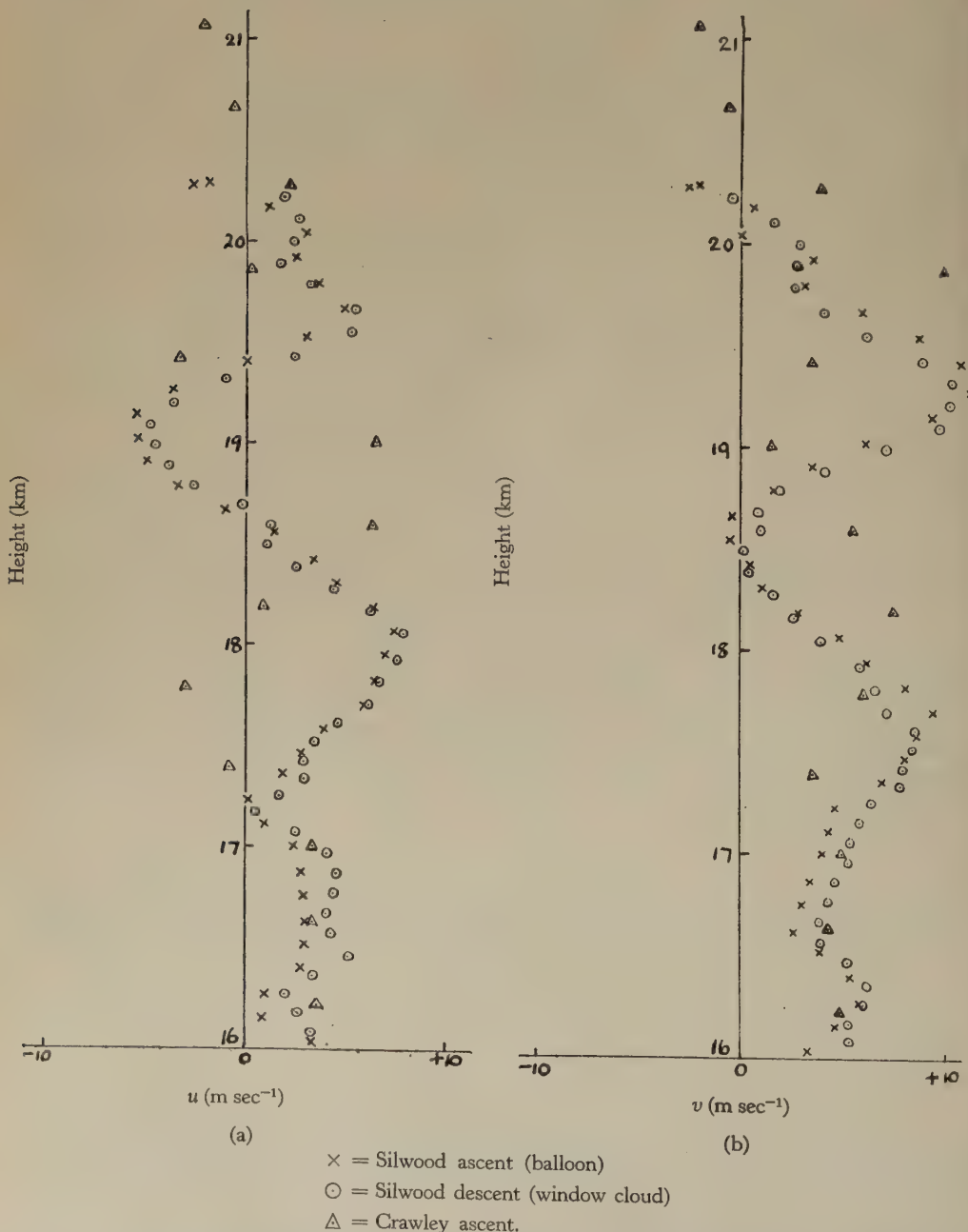

Figure 1. (a) u – eastward and (b) v – northward components of wind, 22 August 1958, during part of the Silwood sounding at 1342 GMT, and the Crawley sounding at 1100 GMT.

Mr. D. C. Thomas : Soundings similar to those described by Mr. Sawyer have been obtained at the Imperial College Field Station at Silwood Park, Ascot using a GL 3 Mk. VII lock-follow radar (a development from the manual-follow radar in general use at upper-air stations in the British Isles). In some of our soundings a cloud of radar-window formed from strips of copper wire was released at burst and tracked during the descent. We found that the cloud behaved effectively as a point source-reflector for approximately 20 minutes after release; during this period its aerodynamic response is similar to that of the balloon used in the sounding. The fall velocity of a window-cloud at 20 km altitude is approximately 6 m sec^{-1} which is nearly equal to the upward velocity of a sounding balloon.

The radar dials were read every 5 seconds using a photographic technique and components of wind (u – eastward, v – northward) computed from readings in groups of 12 with the aid of a digital computer. The method used gave values of u and v at intervals of 20 sec. Fig. 1 shows the results obtained during part of a sounding released at 1342 hr on 22 August 1958. The agreement between the balloon and window-cloud profiles is remarkable, with the difference seldom exceeding 2 m sec^{-1}. 20 minutes elapsed between the balloon and the window cloud readings at 17 km. The maximum fluctuations in u and v are 6 m sec^{-1}, the maximum shear 25 m sec^{-1} km^{-1}, and the vertical wavelength 1·6 km. Fig. 1 also shows the components of wind obtained from the readings taken at one-minute intervals during the routine sounding at 1100 hr at Crawley, 50 km south-east of Silwood. The Silwood and Crawley soundings display considerable agreement with the main features some 500 m higher at Crawley – these soundings are separated by 50 km and over 160 minutes.

A sequence was made of six similar soundings spaced at 3-hr intervals in May 1959 which showed rather smaller fluctuations. It is possible to trace the fluctuations through several soundings in the manner described in the paper and a preliminary analysis suggests that the time-period is between 12 and 24 hours.

I would also like to communicate the following comment to Dr. Murgatroyd's question on the difference shown between the ascent and descent curves in Fig. 2 of Mr. Sawyer's paper.

The standard balloon-sounding rig requires one second to respond to 95 per cent of any change in horizontal velocity, and rises less than 10 m during this time. A reflector falling on a parachute requires approximately 20 seconds to respond equally, and falls between 400 m and 500 m (at 20 km altitude). I have usually found this difference in response distance sufficient to explain the observed differences between ' ascent ' and ' descent ' profiles.

Mr. J. S. Sawyer (*in reply*) : I was very interested to see Mr. Thomas's results which give remarkable confirmation of the Crawley observations using entirely different equipment. His figures comparing the Crawley and Silwood soundings are particularly interesting and suggest a wave-like disturbance with nodal surfaces sloping at a very shallow angle.

In reply to Dr. Murgatroyd, it is difficult to see why such shallow and presumably oscillatory motion should have any material effect on the ozone distribution and I suspect that the ozone variations arise from a different cause.

In reply to Mr. Oddie, I think that we have to imagine superimposed upon the general large-scale wind a system of motion in which air oscillates by converging towards a point at one level from an area of 100 km or more and diverging from a similar point at a level half a kilometre or so lower. Probably 12 to 24 hours later the whole process reverses and the air which was converging diverges and vice versa.

551.510.534

The vertical distribution of atmospheric ozone over Canadian stations from *umkehr* observations

By C. L. MATEER and W. L. GODSON
Meteorological Service of Canada, Toronto

(Read 6 June 1961 at Toronto Conference. See *Q.J.* **86**, p. 512)

Prof. B. W. Boville : The publication of vertical ozone distributions for Resolute, a distinctly arctic station, is noteworthy; the application of such data to a specific stratospheric temperature wave would be revealing.

The comparison of the data for Moosonee to that for Arosa confirms the longitudinal asymmetry of the stratospheric winter temperature and ozone regimes described in the work of the Arctic Meteorology Research Group.

During the current spring the McGill Stratospheric Analysis Project cooperated with Dr. Brewer at Oxford in the hope that ozone sondes might be released at his location during a stratospheric warming period. The final warming wave in the middle stratosphere over western Europe occurred in the period 15-18 March 1961. In a private communication Dr. Brewer has provided ozone data for that period. The total ozone, which was varying from about 300 to 350 m atm-cm prior to the warming, rose to 425 m atm-cm on 19 March. Lower stratospheric amounts varied by about 50 m atm-cm with mixing ratios at 150 mb varying from about 0·3 to 1·5 mg kg^{-1}. Changes were smaller in the vicinity of 60-50 mb. Residual changes indicated by the tops of the sondes and the amounts required to account for the total ozone changes were about 100 m atm-cm above 30 mb with probable ozone mixing ratios at 15 mb varying from 5 to 12 mg kg^{-1}. These measured and suggested changes were reasonably consistent with temperatures varying from − 70C to − 50C at 150 mb and from − 65C to − 45C at 15 mb with lesser changes at 50 mb.

The vertical distributions revealed by these methods tend to verify their broad dynamical dependence whilst posing further problems in detail.

Prof. W. Hitschfeld (*communicated*) : It would be interesting to have the authors' estimate of the relative and absolute precision of their measurements, especially at a time when chemical sondes are producing an increasing quantity of ozone distribution data with high vertical resolution. One notices, for instance, that the authors' information really consists of the ozone amounts in nine 6-km layers, and that these are evidently adjusted so that the sum of the nine amounts equals the total. Are all nine amounts equally reliable, or is the total ozone amount used – as was the practice – to supply essentially one or two of these ?

Dr. W. L. Godson (*in reply*) : I agree with Prof. Boville that Resolute *umkehr* distributions for specific stratospheric strongly-baroclinic temperature waves would be interesting. This has only been possible for March-April 1960, although post-final warming waves have been studied in other years. Relative to the large ozone increase to the annual maximum, associated at high latitudes (but not necessarily in middle latitudes) with the final warming wave of the Arctic middle stratosphere, Moosonee data suggest that the maximum ozone increase occurs in a lower stratosphere layer centred near 130 mb, with a secondary increase in a middle stratosphere layer centred near 35 mb. The corresponding levels at Toronto were 100 and 20 mb.

In reply to Prof. Hitschfeld, all nine layer-amounts were evaluated by a relaxation technique in which the total ozone entered only as the sum of all layer values. It is clear that the construction of a synthetic sounding from the block distribution cannot compete with sonde information, except at upper levels (above 25 km). Inter-comparison tests, such as are planned for Arosa in 1961 and 1962, should give us a clearer picture of the absolute and relative precision of *umkehr* calculations.

551.510.534 : 551.513.2

Total ozone and perturbations in the middle stratosphere

By B. W. BOVILLE and F. K. HARE
McGill University, Montreal

(Read 6 June 1961 at Toronto Conference. See *Q.J.* **87**, p. 490)

Mr. A. H. Gordon : The vertical time section representing upper-air temperatures at Blitzburg in February 1959 shows a pronounced stratospheric low during the latter part of the month. This period was marked by pronounced surface anticyclogensis in N.W. Europe. Is such a temperature distribution in the vertical with its accompanying low ozone content the kind of phenomenon expected to be associated with a markedly anticyclogenic situation ?

Mr. C. E. Thompson : As you know Edmonton, Alberta has been an observation point for total ozone observations for over 10 years. The late Professor Gowan, initiated this study in cooperation with the Meteorological Service, which Service carried it on through the I.G.Y. years until now.

Prof. Gowan, Buckler and myself reported in the Proceedings of the Second National Meeting of the Royal Meteorological Society, Canadian Branch, 18 October, 1956, results regarding ozone changes at Edmonton during the period 1950-1954. In this we looked primarily at the 500 mb level, finding a negative correlation between ozone amount and the height of that pressure surface.

We were looking at the Normand correlation and hence, usually found high ozone in troughs in association with cooling at 500 mb and low ozone on the warm ridges. The relationship of the 500 mb and 30 mb heights is often rather obscure, hence it is difficult to relate our results directly to those achieved by Boville and Hare.

We did notice that summer-type, cold, 500 mb lows in Alberta, sometimes had associated low ozone. This would lead us to wonder whether, in this case, the 500 mb low might have associated 30 mb lows. This would be an interesting point to investigate. Hudson Bay cold lows at 500 mb had very high ozone values at Edmonton, which has some parallel with the warm – high ozone – ridges at 30 mb in Alaska, which sometimes spread into western Canada.

The findings of Boville and Hare indicate major anomalies in the distributions of total ozone as coming from the polar night vortex, hence clarify some of the changes which were inexplicable while looking only at the 500 mb level.

Dr. W. L. Godson : The analytical formulation for ozone amounts in the lower and middle stratosphere appears to be based on an assumed correlation of ozone mixing ratio with potential temperature, with separate regression coefficients for the two layers. The justification for this procedure can hardly be said to have been established by the use of standard *umkehr* distributions which do not necessarily represent a winter average. Actual Moosonee data show rather large departures from these standard curves. It would be interesting to see whether the assumed correlation is in fact valid; the very extensive Arosa data would provide the best test.

It is very difficult to tell to what extent the reported correlations are affected by the assumption of an ozone mixing ratio vs. potential temperature correlation. In particular, it would be interesting to compare the correlation coefficients with those for temperatures alone. It seems peculiar that the lower stratosphere did not account for a significant amount of the residual variance after the strong dependence on 30-mb temperature was removed. Part of the difficulty may lie in the ozone values used; they would only have been actual noon values if a highly reliable observational technique was possible at that hour. Complete values throughout each day are of course available and should be used in correlations of this sort with limited data samples. It is also possible that January 1960 was an anomalous month for the lower stratosphere. A further possibility is that the incorporation of a variable tropopause may have defeated its purpose by implying negative ozone amounts at the base of the layer.

It should be pointed out that ozone in December in the polar regions is not nearly as low as claimed (0·20 atm-cm on the Vigroux scale). Reliable modern observations indicate minima at about the same time as farther south, of a magnitude only slightly less. Amounts have risen appreciably by December, on the average, associated with the increasing dynamic activity of the middle stratosphere. Ozone and temperature are not in phase at this time because infra-red cooling acts to decrease temperatures but has no effect on ozone. For this reason, one should expect seasonal changes in ozone stratospheric-temperature multiple regression coefficients.

Prof. F. K. Hare (*in reply*) : In reply to Mr. Gordon, it must be pointed out that there is no unique relationship between tropospheric anticyclogenesis and the occurrence of very cold stratospheric temperatures. Moreover the anticyclonic phase referred to by Mr. Gordon extended well outside the period of the sharp ozone minimum. Hence the influence of the stratospheric vortex must be assumed paramount at that time.

Mr. Thompson's remarks are of considerable interest. As regards the summer cold lows, there can have been no coupling with middle stratospheric cyclones at that time, since an undisturbed easterly around the polar anticyclone continuously controls the middle stratospheric circulation. Very few tropospheric systems extend higher than the 50 mb surface in summer; hence the ozone effects are presumably wholly due to lower stratospheric activity.

Dr. Godson's strictures are to the point; but they seem to reinforce rather than weaken our arguments. We did not have access to the complete ozone readings throughout the day at the time we prepared the paper; our high correlation values were thus obtained in spite of this possible weakness. We are aware of the fact that Moosonee's actual ozone profiles often differ from the standard *umkehr* profiles; yet the correlation obtained is still highly significant. It would of course, be futile to use Arosa data for a test, since in most winters Arosa lies almost permanently outside the area of the main polar-night westerly circulation.

Proceedings at meetings of the Society

Ordinary Meeting, 17 May 1961

The following candidates were declared to have been elected that day by the Council :

Fellows

Balogun, Elisha Ekumdayo, Meteorological Services Headquarters, Lagos, Nigeria, West Africa.
Browning, Keith Anthony, B.Sc., A.R.C.S., Department of Meteorology, Imperial College Field Station, Silwood Park Farm, Nr. Ascot, Berks. (*transfer from Student Membership*).
Golde, Rudolf Heinrich, Ph.D., D.Sc., 274 Salmon Street, London, N.W.9.
Griffith, Miriam Violet, B.Sc., F.Inst.P., 89 Deanhill Court, East Sheen, London, S.W.14.
Griggs, Michael, B.A., The Clarendon Laboratory, Parks Road, Oxford (*transfer from Student Membership*).
Hamilton, Paul Marshall, c/o Stormy Weather Group, Physics Department, McGill University, Montreal 2, Canada (*transfer from Student Membership*).
Hancocks, William Malcolm Bramwell, B.A., 12 Devonshire Road, Cambridge.
King, Robin Humphrey, B.A., c/o Meteorological Office Training School, Stanmore, Middx.
McFarlane, David, Meteorological Office, Royal Air Force, Nicosia, B.F.P.O. 53.
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Shomefun, Ganiyu Akanni, Meteorological Service Headquarters, Lagos, Nigeria, West Africa.
Stubbs, Martin William, B.Sc., 61 Allangate Road, Liverpool 19 (*transfer from Student Membership*).
Wardhaugh, Keith Grant, B.Sc., 21 Bank Steps, Silver Street, Durham (*transfer from Student Membership*).
Whittall, Leonard William, Spring Cottage, Crazies Hill, Wargrave, Berks.
Winters, David, 66 Hawthorn Way, Chesterton, Cambridge.

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Assil, Mohamed, Hilal Hotel, Herika Square, Hussayni Street, Damascus, U.A.R.
Cormenzana, Enrique, c/o World Meteorological Organization, 41 Avenue Giuseppe Motta, Geneva, Switzerland.
Vadhanapanich, Capt. Charoen, B.S., M.S., Ph.D., R.T.N., Meteorological Department, Bangkapi, Bangkok, Thailand.

Student Members

Abbott, Mary, 46 Blackacre Road, Theydon Bois, Epping, Essex.
Cox, Howard John, 184 Hillmorton Road, Rugby, Warwicks.
Green, Neil Peddie, 72 William Way, Letchworth, Herts.
Greenwood, David Michael, 305 Mutton Lane, Potters Bar, Middx.
Mackay, Ian Robert, Ridgewood, Queen Hoo Lane, Tewin, Welwyn, Herts.
Rodgers, Douglas, c/o Gonville & Caius College, Cambridge.
Williamson, Edward James, The Clarendon Laboratory, Parks Road, Oxford.

Canadian Branch

Fellows

Cork, Howard, F., B.Sc., c/o 15 Tunbridge Crescent, Etobicoke, Ontario.
Cudbird, Beverley Swann, B.A., 7 Shaver Court, Islington, Ontario.
Davies, Davied, M.A., 160 Dorval Avenue, Apt. 18, Dorval, P.Q.
East, Rev. Father Conrad, B.Sc., M.A., College Jean-de-Brebeuf, 3,200 Cote St. Catherine, Montreal 26.
Gilmour, William David, B.Sc., Aviation Forecast Office, R.C.A.F. Station, Winnipeg, West Manitoba.

Smith, Wendell R., M.A., 123 Young Boulevard, Toronto 12, Ontario.
Sporns, Ulrich, 21 Hapley Drive, Scarborough, Ontario.
Tripp, Richard B., Pacific Oceanographic Group, Biological Station, Nanaimo, British Columbia.

Special General Meeting, 21 June 1961

A Special General Meeting of Fellows was held to elect the Officers and Council for the year 1961-62.

Fellows had been circulated on 19 April with the list of nominations by Council for Officers and Council for the year 1961-62. No other nominations having since been received the persons so nominated were declared to be elected unopposed in accordance with By-law 53.

The Officers and Council for 1961-62 were therefore elected as follows :

President

Howard Latimer Penman, M.Sc., Ph.D., F.Inst.P.

Vice-Presidents

Arthur George Forsdyke, A.R.C.S., Ph.D.
Patrick John Meade, O.B.E., B.Sc., A.R.C.S.
John Stanley Sawyer, M.A.
James Martin Stagg, C.B., O.B.E., M.A., D.Sc.

Vice-President for Canada

Frederick Kenneth Hare, Ph.D.

Vice-President for Scotland

Robert Cranna, M.A., B.Sc.

Treasurer

Brian Cecil Vernon Oddie, B.Sc.

Secretaries

George Arthur Bull, B.Sc., M.I.Inf.Sc. (Librarian)
Ronald Frith, O.B.E., M.A., Ph.D.
Frank Pasquill, D.Sc. (Editor)
Richard Segar Scorer, M.A., Ph.D.

Councillors

John Alan Chalmers, M.A., Ph.D., F.Inst.P.
Tony John Chandler, M.Sc.
Ronald Harry Clements, M.A.
John Robert Dark Francis, B.Sc. (Eng.), A.I.C.E.
William Angus Grinsted, C.B.E.
John Theodore Houghton, M.A., D.Phil.
Ernest Knighting, B.Sc.
Arthur Leonard Maidens, B.Sc.
John Lennox Monteith, Ph.D.
Paul Robert Owen, B.Sc., A.F.R.Ae.S.
John Russell Thorp, O.B.E., M.A., R.N.
Gilbert Brian Tucker, Ph.D.

Ordinary Meeting, 21 June 1961

The following candidates were declared to have been elected that day by Council :

FELLOWS

HAWES, Charles Godfrey, B.Sc., 14 Northway, Morden, Surrey.
MELLENFIELD, Alan Keith, 281 Friern Barnet Lane, Whetstone, London, N.20.
PICKERING, Kenneth, 18 Wycherley Crescent, Barnet, Herts.

STUDENT MEMBERS

KING-HELE, John Andrew, 7 Auckland Road West, Southsea, Hants.
MORRIS, Ruth Elizabeth, 18 St. Malo Road, Wigan, Lancs.

CANADIAN BRANCH

FELLOW

MAYBANK, John, B.Sc., M.Sc., Ph.D., Ralston, Alberta,

The following papers were read and discussed :

1. ' Precipitation over the North Atlantic Ocean ' by G. B. Tucker (*Q.J.*, **87**, p. 147).
2. ' Some experiments in the objective analysis of contour charts ' by G. A. Corby (*Q.J.*, **87**, p. 34).
3. ' Quasi-periodic wind variations with height in the lower stratosphere ' by J. S. Sawyer (*Q.J.*, **87**, p. 24).

Summer Visit, 5 July 1961

The Summer Visit of the Society was paid this year to the Warren Spring Laboratory at Stevenage, Hertfordshire, the newest of the 15 research stations of the Department of Scientific and Industrial Research. By the kind invitation of Mr. S. H. Clarke, C.B.E., the Director of the Laboratory, some 40 Fellows and their friends took tea with him and his senior staff members, and the objects and scope of the Laboratory were described in a most interesting talk.

The party then broke up into four groups and were conducted on a tour of the departments in which exhibits were to be seen ranging from a demonstration of an ingenious air-pollution collector and a portable sulphur-dioxide sampler to experiments in the chemical-engineering section dealing with the absorption of noxious gases from industrial effluents, and also to the pilot-plant buildings where diesel exhaust fumes were being measured.

The tour impressively evidenced the wide range of important research being carried out and when the party reassembled before departure, the time allotted for questioning the demonstrators was all too short.

In the unavoidable absence of the President, Dr. Scorer proposed the vote of thanks for a most interesting and enjoyable afternoon which was warmly endorsed by all the visitors. A fuller account of the visit will appear in *Weather*.

Summer Meeting, Exeter, 12-14 July 1961

The eighth two-day Summer Meeting of the Society was held at the University of Exeter, by kind permission of the Vice-Chancellor, Dr. J. W. Cook, F.R.S., and the University Senate.

The meeting was formally opened on the evening of 12 July at a reception given by the university authorities in Mardon Hall, where the visiting members of the Society were most comfortably accommodated. Professor G. K. T. Conn, M.A., Ph.D., who was accompanied by Mrs. Conn and Miss Richardson the House Manager, welcomed the visitors in a short speech to which Dr. J. M. Stagg, President of the Society, replied.

At the first symposium, on ' Winds and circulations in the upper atmosphere,' Dr. Stagg took the Chair. Speakers included D. R. J. Murgatroyd who read his and Mr. F. Singleton's paper ' Possible meridional circulations in the stratosphere and mesosphere.' After discussion, Dr. J. S. Greenhow read his and Dr. E. L. Neufeld's paper on ' Wind at 80-100 km as revealed by radar reflections from meteor trails.' Discussion followed and Mr. R. A. Ebdon then presented his paper on ' Fluctuations in tropical stratospheric winds,' and discussion concluded the session.

The Chair at the second day's symposium on ' Medium-scale diffusion in the atmosphere ' was taken by Dr. H. L. Penman. Dr. F. Pasquill read his paper on ' Some observed properties of medium-scale diffusion in the atmosphere.' After discussion, Dr. J. M. Hirst read his paper on ' The long-range travel of fungus spores.' Following discussion, Dr. R. S. Scorer presented his paper on ' Mechanisms of atmospheric diffusion,' and the session concluded with discussion.

On Thursday afternoon, the City Engineer and Surveyor, Mr. J. Brierley, gave an illustrated talk on the causes and effects of floods in the area. This was followed by a tour of the outlying neighbourhood and sections of Exeter which has been so seriously affected by floods in the recent past. Apart from the interest, the afternoon was most enjoyable.

On Friday afternoon, inclement weather unfortunately impaired full enjoyment of the coach tour of Dartmoor which had been arranged, although it was a pleasant excursion for those who ventured.

In the later afternoon, several members availed themselves of the generous offer of Mr. Roderick Ross, M.A., Secretary of the university, to accompany them on a personally conducted tour of the new university buildings, which all agreed were some of the finest they had seen.

Council wishes to record its sincere thanks to Professor B. J. Mason who acted as co-ordinator, and to all whose efforts assured the success of the meeting.

A full account of the meeting will appear in *Weather*.

Reviews

Meteorology for glider pilots. By C. E. Wallington. London (John Murray), 1961. Pp. xii, 284; 116 Figures; Tables and 22 photographs, 25s.

A good book on this topic has not been written before because there was no one with sufficient experience to do so. Since 1953, Mr. Wallington has been more closely associated with forecasting for British glider pilots than anyone else and he has a keen appreciation of what they need and what they can take. But since the kind of weather which glider pilots are interested in is, in the large, just that which concerns us all, it is a valuable book for anyone. Although written for a special audience it possesses a liveliness and excitement at new things discovered, which is not usually found in books on bread-and-butter meteorology for the ordinary reader.

It is no easy task to explain to a non-specialist readership – keen though the gliding community is – without some simplification and therefore misrepresentation. For example, the halo (p. 51) often, but by no means usually, seen ahead of warm fronts, is found as commonly elsewhere (and the picture given was actually taken under false cirrus from cumulonimbus); on p. 45 it is suggested that the material collected in a deposit gauge is primarily the cause of bad visibility (but this is normally due to much smaller particles); the standard theory (p. 33) that turbulence produced over rough flat terrain can alone cause cloud at several thousand feet just won't bear close scrutiny; Berg winds (p. 114) are not just katabatic – they are too warm, and can we really believe that ana-flow produces kata-flow on the opposite side of a pass (p. 114) ? Ana-winds are not just kata-winds in reverse as implied in the diagrams – they are much deeper. The standard explanation of the Föhn effect (precipitation on mountains) won't always do and the diagram (p. 137) showing a 10,000 ft mountain is scarcely representative of Britain where the effect is very widespread. On p. 178 it is suggested that ' chaotic motion ' is responsible for the collision of ice particles of different temperature in a Cb, and on p. 182 the author cannot get away from the idea that, because the air in a cloud is statically unstable, down-draughts are almost spontaneous. That old myth that ' more recent researches ' have shown the Bergeron theory to be inadequate is revived : actually warm rain was well known to forecasters a generation ago. The dynamical explanation of a thunderstorm high, on p. 183-4, is inadequate and (p. 198) there *are* reasons why a cu base should be concave. The author admits that his spring analogy for the stratification of the atmosphere has its faults – I agree ! He might have drawn at least one rotor with orderly flow because these are not all turbulent. This reviewer did not invent ' fumulus ' nor (p. 46) did he apply it to a *layer* of smoke. Nor do I like the classification of clouds, which is archaic (though admittedly quite orthodox), because wave clouds – so important to the glider pilot – are omitted.

If these complaints were about things which the author had written off his own bat they would not be made : the trouble is that he is perpetrating the ' official ' view and the errors will be quoted and quoted again for another generation. The author is criticised for not being as independent as we would wish him to be : nevertheless this is essentially a new book with a welcome liveliness, to be heartily recommended.

Anyone who has the essential physics can start without any background reading and get farther in ' Wally's ' 284 pages than in any other book of comparable size. It is about the phenomena we see and feel, and the charts are brought to life with drawings of cross-sections placed upright upon them. The writing is directed at people who will be out of doors – which is where the weather is, and each phenomenon is illustrated by charts and diagrams of actual instances, with magnitudes given. This is particularly true of the sea breeze (which we might all think we know from the geography books), for we meet the sea-breeze front (described by Peters 25 years ago but almost ignored since) on the second page of Chapter 11. With the tephigram we have the ' golden ' numbers giving the effects of insolation and there is a mine of information at the end of the book about broadcast weather forecasts, average sea and land temperatures, telephone numbers of forecasting offices, conversion factors and useful references.

Both physics and geography teachers could well recommend it for their school library; naturalists, air-line passengers, sportsmen, and anyone interested in getting the most out of the forecast, including glider pilots, will find Wallington's easy approach an encouragement to further reading.

R.S.S.

Handbook of aviation meteorology. Air Ministry, Meteorological Office, London (H.M. Stationery Office), 1960. Pp. xvi, 404; 147 Figures; 12 Tables; xxvi Plates. 25s.

Every meteorologist who has provided service to aviation at any time during his career must surely be familiar with the renowned *Meteorology for aviators* written by Dr. R. C. Sutcliffe and first published in 1940. He must also be well aware of its universal acceptance by pilots and navigators as an authentic primer of aviation meteorology. With the rapid development of aviation during the last twenty years resulting particularly in regular flights at ever increasing altitudes there has been a need for some time for an up-to-date ' Sutcliffe ' and, as stated in the preface, *Handbook of aviation meteorology* was prepared by the Meteorological Office as a successor to *Meteorology for aviators*, which is now out of print.

The compilation of the book is attributed largely to Crossley from material built up by him whilst engaged in setting and marking examination papers for civil air-navigation licences and it is admirably suitable for use by air crew as a standard textbook on meteorology. Indeed, it is better to treat this publication as a textbook giving a physical account of elementary meteorological theory and of meteorological phenomena rather than as a handbook wherein one might hope to find information of direct application to day-to-day flight operations.

The book is divided into five parts and the material is well presented and amply illustrated. Part I deals with physical principles and their applications to aviation and one chapter is devoted to characteristics of high-altitude flight. Part II describes how surface and upper-air meteorological observations are made, coded and charted. The section on cloud is treated in detail and includes 26 Plates of cloud photographs, some of which are taken from the *International Cloud Atlas*. Part III gives an outline of the principles of weather forecasting. Some typical synoptic situations affecting weather over the U.K., and in one case over the Mediterranean, are discussed from sequences of weather charts over a few selected periods. Part IV on meteorological organization occupies over five pages and appears to contribute little to the value of the book. Certain sections, particularly those concerning in-flight procedures, are out-of-date and require amendment. Part V, on climatology, includes a chapter on the general circulation and world climate and a chapter on aviation climatology of air routes. The latter gives a good account of the significant features of the weather over most of the trunk routes including a number of airfields, but omits West Africa and the Pacific Ocean.

On the whole the book *is* a worthy successor and it will be of value not only to aviators but to anyone interested in meteorology.

E.C.

Meteorology and climatology for sixth-forms. By E. S. Gates. London (George G. Harrap), 1st Ed., 1961. Pp. 203; 9 Plates; 113 Figures; 18 Tables. 13s. 6d.

Can meteorology and climatology be profitably taught above middle-school level without resort to the solution of mathematical problems ? Opinions differ.

Mr. Gates has made a creditable attempt to explain the physical basis of these subjects with a minimum of mathematical stress. The author has the advantage of having taught geography at sixth-form level and I should imagine, sixth-form physics as well. His aim has been to pursue what, for the average sixth-form pupil, must be a fairly rigorous road, without pauses for rest in those fascinating yet delusory side paths characteristic of many popular books about the atmosphere.

Sometimes, indeed, the simplification seems to have been carried too far. In Fig. 97 there is a map of the summer flow of air into Asia which seems to ignore the circulation over N.W. India and the effect of the mountain barrier, though there is a brief reference to the latter in the text. There is a curious mixture of precision and vagueness in the statement on p. 179 that ' approximately 59 per cent ' (of Marine West European precipitation) ' occurs in the winter.' It is disappointing to find such a wretched bibliography in a book which is likely to whet the appetite of young persons with a taste for science.

These, however, are minor criticisms and should not be allowed to overshadow the general excellence of this book. Mr. Gates, according to his preface, has written for geographers but there is much in his book which will be useful reading for the forward-looking young physicist anxious to widen his background. For such a reader there are excellent sections on the general law of gases in the atmosphere, heat balance, geostrophic winds, humidity, adiabatic cooling, atmospheric optics and electricity.

This book deserves a wide public and a place in every library where scientific knowledge may be sought.

J.B.R.

Antarctic meteorology. Proceedings of the symposium held in Melbourne, February 1959. Oxford (Pergamon Press), 1960. Pp. xviii, 483; 3 Plates; Figures and Tables. 5 gns.

The practice of collecting papers presented at a scientific symposium and printing them in a relatively expensive publication is not one which commends itself to scientists in general, and to meteorologists in particular. A mitigating factor in the case of *Antarctic meteorology* is that this is probably one of the largest and most varied collections of papers in this subject to be assembled in one volume. However a general impression obtained from the contents is that many of the contributors never intended their papers for such an elaborate publication. The object of the symposium was to air ideas and to present and discuss some of the observations made during the International Geophysical Year as an aid to further analyses. There can hardly have been time for much detailed analysis and although some new and interesting work is presented, many of the papers are intended to do no more than present the data, whilst others are reports on the progress of preliminary surveys. Several reports end by saying that the material in them should be considered tentative and may be subject to revision in the light of further data.

Some papers present nearly ' raw data,' others are theoretical and synoptic analyses. The mixture of these two types of contribution in a symposium is excellent because of the mutual effect the authors have on the work of each other; in a scientific publication, however, they are hardly complementary.

The Symposium was obviously a great success, with six sessions, one on each of the following topics : Local effects in the Antarctic; Synoptic analyses and forecasting; Snow and ice characteristics; Heat and mass exchanges; Climatological aspects.

It would be invidious to single out some of the contributions, but the papers of the Australian and the U.S.S.R. participants must be mentioned for their comprehensive and varied approach to the general problem. One by R. H. Clarke on ' An attempt to deduce the effect of Antarctica on the temperature of Middle Latitudes by a controlled-model experiment ' deserves special mention. It describes the application of the well-known dish-pan technique and results suggest that differences between the dynamical characteristics of the Northern

and Southern Hemispheres can be largely attributed to differences in surface roughness. Both the technique and the result aroused much interest and one hopes to hear soon of further experiments along these lines.

The tome is interesting mainly because it outlines the various methods (statistical, synoptic, theoretical and experimental) which are being employed in studying the problems posed by Antarctic meteorology.

Apart from the reservations given in my first two paragraphs, the papers, and particularly the references given in this book, may well be very useful to those working in this field. The production – print, diagrams, paper and binding – are excellent.

G.B.T.

Handbuch der Aerologie. Ed. by W. Hesse. Leipzig (Akademische Verlags. Geest I Portig.), 1961. Pp. vii, 897; 455 Figures; 91 Tables. 80 D.M.

The editor, W. Hesse, writes in the Introduction that the icea of producing a textbook of aerology was due to Max Robitzch whose work in preparing such a book was curtailed by his death. The idea was taken up by a group of meteorologists under Hesse's editorship and led to the volume now reviewed. *Handbuch der Aerologie*, aims at setting down the methods of observation in the free atmosphere, the reduction and interpretation of these observations and their use in both forecasting and climatology. Sixteen authors, each a well-known specialist in the field in which he writes, have contributed, and since the range is great it is difficult for one meteorologist to write a balanced critical review. However, since the book is presumably not aimed at other specialists but written for meteorologists in general, it must be judged by its appeal to them.

Each chapter of the book contains some 50 to 100 pages and is supported by a bibliography of 100 to 200 entries, so that the first impression is one of completeness. This impression is strengthened by the general principle of writing the chapters on an historical basis, dealing first with quite early work, and in fact the handbook is also an history of aerology from the earliest days up to the present day. It is certainly up to date, for several references are to papers published in 1960. The completeness makes it perhaps unlike any other meteorological text, for we find such dissimilar subjects as methods of launching a balloon, radiosonde circuitry, theoretical considerations of measurements made by rockets and methods of constructing climatological statistics. It is, therefore, a true handbook, not heavily weighted in one particular sphere and no meteorologist will be professionally interested in the whole of the book.

The first three-quarters of the book is devoted to the making of measurements in the free atmosphere. Dr. Hesse gives first a short account of the history of aerology and follows with a chapter on the thermodynamics of the atmosphere, including the use of upper-air diagrams. This is a very comprehensive account of subject matter which is difficult to make interesting. Then follow two contributions on the ' classical ' methods of making upper-air observations over land and sea. The historical information about kite and balloon flying is delightfully interesting and the fearlessness of early investigators is marked by Fig. 41 and others; their ingenuity is also marked. Information is given, for the first time as far as this reviewer knows, of the methods adopted in Germany to fly instruments at a fixed point in space from a tethered balloon in the immediate pre-war period, c. 1937.

There are 400 pages concerned with more recent methods of making observations of many different sorts in the free atmosphere, ranging through the use of gliders, aircraft and rockets as well as free balloons as the instrument platform. The descriptions of instruments and their characteristics are full, whilst at the same time there is ample reportage and discussion of measurements (e.g., in the chapters on the measurement of ozone, and on atmospheric turbulence).

The two final chapters deal with the use of the measurements, firstly in synoptic meteorology and secondly in climatology, and of course these two chapters deal with only some of the observations that are made, since synoptic meteorology does not, as yet, use measurements of, say, ozone, and some of the observations are so recent in concept of measurement that their climatology is not practicable.

This then is an authoritative encyclopaedic book, useful to one wishing to find out technical data concerning different radiosondes, to another wishing to refer to sound propagation in the atmosphere, or to another dealing with electrical effects in the atmosphere and so on. It is beautifully produced with fine line drawings and many photographs; the mathematics are well displayed and it has a comprehensive index. It should be in every meteorological library.

E.K.

Einführung in die Klimatologie. By R. Scherhag. Braunschweig (G. Westermann Verlag), 1st. Ed., 1960. Pp. 128; 9 Figures; 12 Tables; 11 Charts. 8·80 D.M.

This book is one in the series *Das Geographische Seminar* and, as its title implies, covers a wide field but does not attempt to deal in detail with any of the topics discussed. The author has clearly kept his geographical readers in mind and, apart from stating a few formulae, he has avoided any mathematical treatment. The book is divided into sections, of which the four most important are concerned with the fundamental laws of meteorology; planetary circulation; pressure- and wind-belts; and climatic classification. These account for some three-quarters of the book and the remainder comprises seven shorter sections on climatic variations and changes; microclimatology; the influence of climate on man; the effect of human activity on the climate; the scientific methods of climatology (which deals with diurnal variations and with simple statistical ideas); and, finally, the weather chart.

It is somewhat surprising that the author has made so little use of diagrams, of which there are only nine in the body of the text. The lack of a diagram is particularly noticeable in the sections on pressure – and wind-belts, and climatic classification. There are, however, eleven charts with the book but, although some of them are briefly referred to in the text, there is little integration between charts and text. Some charts are unexpected in a book of this type; for example, the mean contours for the 50 mb and 100 mb surfaces in January and July.

The book is a paper-back but is well produced and convenient in size. One can therefore hope that it will be available to many students in Germany and that some will be led to follow up this introduction to the wide range of topics which Professor Scherhag has provided.

W.H.H.

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Translations of Russian meteorological literature

To assist in making Russian meteorological literature more readily available the Society, in 1958, proposed to certain other meteorological institutions that each should undertake the translation in whole or in part of Russian meteorological journals, the translations to be distributed by exchange between participants. To avoid duplication the Society undertook to act as coordinator.

For its own part in this scheme the Society proposed to translate the journal *Works of the Central Aerological Observatory of the U.S.S.R.* Translations of the first two papers are now ready. They are :

Works of the Central Aerological Observatory, Pt. 21, 1958.

' Investigation of the variability and diurnal change of wind in the atmosphere,' by V. D. Reshetvo, pp. 3-51.

' Investigation of the pulses of the horizontal component of wind up to a height of 5 km,' by V. A. Devyatova, R. V. Pyatyshev, R. O. Tydelskaya and I. A. Cherenkova, pp. 52-175.

The translations are complete with tables and diagrams.

Some 50 subscribers to the *Quarterly Journal* have expressed a wish to buy these translations, and, accordingly, copies have been duplicated. A number of copies are still available, at the price of 3 guineas, and may be had on application to the Assistant Secretary.

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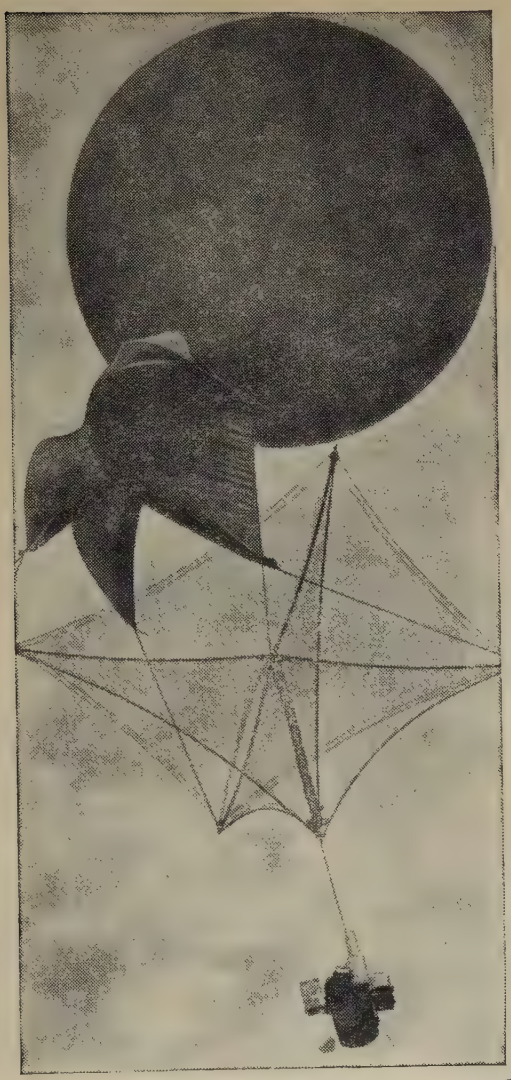

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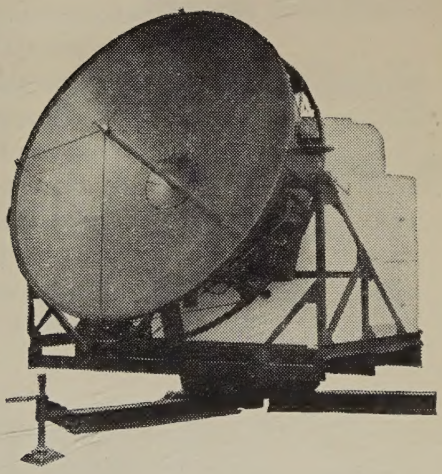

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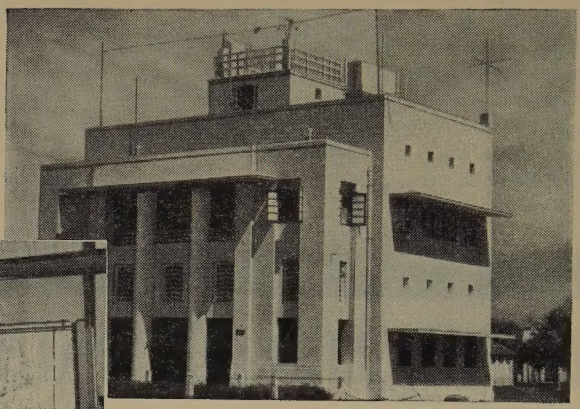

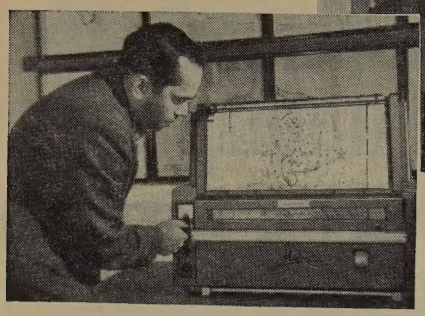

In the book "Arthastra" by Chanakya, founder of the Maurya dynasty in India, it appears that the Indians of the fourth century B.C. studied meteorology. The book states that the quantity of rain varied between 13½ dronas and 23 dronas. These days, rainfall is not measured in dronas, but India is still one of the most progressive countries as regards instrumental equipment.
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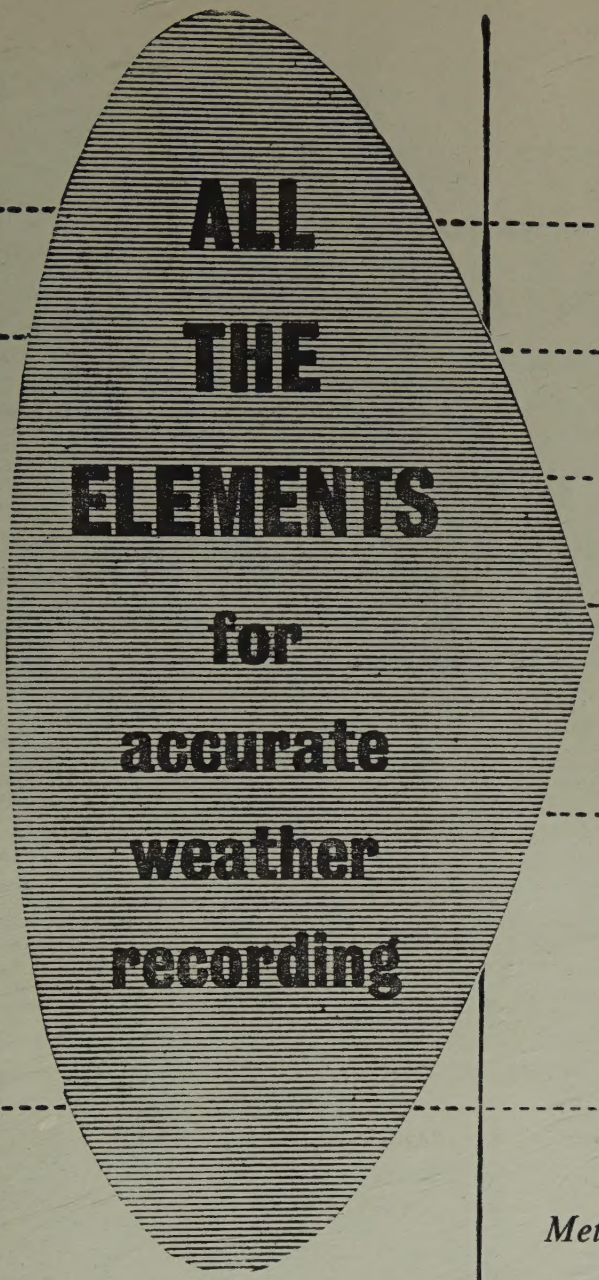

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THE NAPIER SHAW MEMORIAL PRIZE

Announcement of third competition

The Royal Meteorological Society announces the third competition for the Napier Shaw Prize. The competition will be open to anyone without restriction of nationality. The prize of £100 is offered on this occasion for an original essay on '**The Upper Atmosphere.**'

> For the purpose of this competition the upper atmosphere is regarded as the atmosphere above the tropopause and below the exosphere. Essays on the dynamics, constitution or physics of the atmosphere in this region, or in any part of it, will be accepted, but an essay on purely electrical properties of the ionosphere is not sought. Preference will be given to essays relating two or more aspects of the state of the upper atmosphere. Essays on the inter-relations between the upper atmosphere and the troposphere will also be welcome.

The essay, which must be in English, may include material which the candidate has already published. Essays may be submitted jointly in the names of more than one author.

Candidates are required to submit three copies of their essay, which must be printed or typewritten. In the latter case, only one side of the paper should be used, double spacing should be employed and two of the copies may be good carbon copies. Unless the essay is printed, any mathematics should be written in carefully by hand. Although the scale of the essay will be considerably greater than that of an ordinary paper, candidates are recommended to consult, for general advice, the pamphlet, 'Preparation of papers for the Quarterly Journal.' This may be obtained, free of charge, on application to the Assistant Secretary.

It is required that any essay for which a prize is awarded shall be offered to the Society for publication.

All entries for the competition should be addressed to the Assistant Secretary, Royal Meteorological Society, 49 Cromwell Road, London, S.W.7 and marked clearly 'Napier Shaw Prize.'

Essays for the competition must be received at the Society's Offices *not later* than 1 January 1962.

Printed in Great Britain in Horley Series by The Salisbury Press Ltd., Publicity House, Wilton Road, Salisbury, Wilts. (30194)